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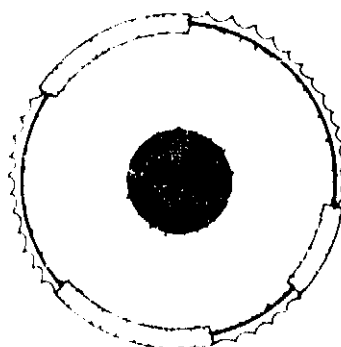
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WORKSHOP ON

THE EARLY EARTH: THE INTERVAL FROM ACCRETION TO THE OLDER ARCHEAN



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WORKSHOP ON
THE EARLY EARTH:
THE INTERVAL FROM ACCRETION
TO THE OLDER ARCHEAN

Edited by
Kevin Burke
and
Lewis D. Ashwal

A Lunar and Planetary Institute Workshop
April 23-24, 1984

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Contents

Introduction	i
Program	iii
Summaries of Technical Sessions and Discussions	vi
Abstracts	
<i>The Relation Between the Age of the Subducting Slab and the Recycling of Sediments into the Mantle</i> D. Abbott and S. Hoffman	1
<i>Constraints on Early Atmosphere from Planetary Accretion Processes</i> G. Arthenius	4
<i>The Composition of the Primitive Atmosphere and the Synthesis of Organic Compounds on the Early Earth</i> J. L. Bada and S. L. Miller	8
<i>Conditions During Formation of the Earth</i> A. G. W. Cameron	11
<i>Early Evolution of the Crust-Mantle System</i> K. C. Condie	14
<i>Was Core Formation Violent Enough to Homogenize the Early Mantle?</i> S. A. Coopernan and W. M. Kaula	17
<i>The Problem of the Earth's CO₂ Content and the Iron Core</i> G. W. DeVore	20
<i>Considerations of Large Scale Impact and the Early Earth</i> R. A. F. Grieve and E. M. Parmentier	23

<i>Hydrothermal Alteration in the 3.5 b.y. old Onverwacht Group of South Africa</i> R. Hart and M. J. de Wit	25
<i>The Degassing History of the Earth; Noble Gas Studies of Archean Cherts and Zero Age Glassy Submarine Basalts</i> R. Hart and L. Hogan	29
<i>The 3.5 b.y. old Onverwacht Group: A Remnant of Ancient Oceanic Crust</i> S. Hoffman	31
<i>Submarine Hydrothermal Environments as Sites for the Origin and Evolution of Life</i> S. Hoffman and J. Baross	34
<i>Noble Gases and the Early History of the Earth: Inappropriate Paradigms and Assumptions Inhibit Research and Communication</i> G. R. Huss and E. C. Alexander, Jr.	37
<i>Constraints on the Development of the Early Continental Crust from Isotopic Data</i> S. B. Jacobsen	39
<i>Anorthositic Oceanic Crust in the Archean Earth</i> E. Jagoutz, J. B. Dawson, S. Hoernes, B. Spettel, and H. Wänke	40
<i>A Model for Core Formation in the Early Earth</i> J. H. Jones and M. J. Drake	42
<i>Implications for the Earth of the Early Dynamical Environment</i> W. M. Kaula and S. A. Cooperman	45
<i>A Review of Tectonic Aspects of the Limpopo Belt and Other Archean High-Grade Gneissic Terranes</i> W. S. F. Kidd	48
<i>A Review of Noble Gas Geochemistry in Relation to Early Earth History</i> M. D. Kurz	50
<i>Archean Sedimentary Systems and Crustal Evolution</i> D. R. Lowe	51

<i>The Origin of Continental Crust: Outlines of a General Theory</i> P. D. Lowman, Jr	54
<i>The Aravalli Sequence of Rajasthan, India: A Precambrian Continental Margin?</i> J. D. Macdougall, R. Willis, G. W. Lugmair, A. B. Roy, and K. Gopalan	57
<i>Carbon and Nitrogen in the Deep Earth</i> D. P. Matthey, L. P. Carr, R. H. Carr, I. P. Wright, and C. T. Pillinger	59
<i>Crustal Radiogenic Heat Production and the Selective Survival of Ancient Continental Crust</i> P. Morgan	62
<i>Water Loss from Venus: Implications for the Earth's Early Atmosphere</i> S. M. Richardson, J. B. Pollack, and R. T. Reynolds	65
<i>Chemical Evolution of the Earth—Equilibrium or Disequilibrium Process?</i> M. Sato	68
<i>Large Arcuate Scars: A Geological Legacy of the Earth's Accretionary Past</i> J. M. Saul	71
<i>Lunar and Martian Impact Basins: Exposed Records of Terrestrial Bombardment?</i> P. H. Schultz	74
<i>Thermal, Dynamic and Compositional Aspects of the Core-Forming Earth</i> D. J. Stevenson	76
<i>Earth's Primordial Differentiation, and Its After-effect.</i> P. H. Warren	79
<i>Chemical Constraints on the Evolution of Archaean Continental Crust</i> B. L. Weaver and J. Tarney	82
<i>Dating Archaean Zircon by Ion Microprobe—New Light on an Old Problem</i> I. S. Williams, P. D. Kinny, L. P. Black, W. Compston, D. O. Froude, and T. R. Ireland	84
Registered Attendees	87

Introduction

Fifteen years of research following the first return of rocks from the Moon has enabled planetary scientists to develop a framework in which to consider the mechanisms whereby planetary bodies are constructed. Concurrently, scientists have been studying the oldest available terrestrial rocks to try to understand how the Earth operated in the most distant geologic past. The Earth is a highly dynamic planet, however, and the record of events prior to about 3.8 b.y. years ago has been almost completely obliterated. We are thus left with a gap of about 800 million years between the time of Earth's accretion and that of the oldest preserved rocks.

The Early Earth Workshop was convened as an attempt to understand what may or may not have occurred during that period and to review current understanding of the Earth's formation and the formation of the older Archean rocks. One goal of this workshop was to determine what theoretical, observational, and/or experimental studies can best constrain the character of the Earth in this important interval of time.

Fifty-five scientists with backgrounds in a broad range of disciplines, including planetary accretion, isotope geochemistry, thermal history modeling, and Archean geology, met at LPI in Houston in April, 1984 to discuss and speculate on this fascinating interval of the Earth's history. This volume constitutes a report of the workshop, consisting of abstracts of the presented papers and brief summaries prepared by session coordinators. We hope that this summary of currently available facts, observations, and hypotheses may prove stimulating to those who feel ready to influence our concepts about earliest Earth history and evolution.

The workshop constitutes part of the Institute's Early Crustal Genesis (ECG) program. Other workshops and meetings in this program for which Technical Reports are available or soon to be published include the Workshop on Magmatic Processes (Tech. Rpt. 82-01), the Conference on Planetary Volatiles (Tech. Rpt. 83-01), the Workshop on A Cross Section of Archean Crust (LPI Tech. Rpt. 83-03), and the recently held Conference on the Origin of the Moon (which will be published in book form by the LPI in early 1986).

Program

Monday, April 23
9:15 a.m.-12:30 p.m.

Accretion of the Earth: Processes Accompanying and Immediately Following the Accretion, Including Core Formation

George Wetherill, Coordinator

Conditions During Formation of the Earth
 A. G. W. Cameron

Was Core Formation Violent Enough to Homogenize the Early Mantle?
 S. A. Cooperman and W. M. Kaula

A Model for Core Formation in the Early Earth
 J. H. Jones and M. J. Drake

Implications for the Earth of the Early Dynamical Environment
 W. M. Kaula and S. A. Cooperman

Chemical Evolution of the Earth—Equilibrium or Disequilibrium Process?
 M. Sato

Thermal, Dynamic, and Compositional Aspects of the Core-Forming Earth
 D. J. Stevenson

Monday, April 23
2:00-3:30 p.m.

Impact Records and Other Information From Planets and the Moon Relevant to Early Earth History

Richard A. F. Grieve, Coordinator

Considerations of Large Scale Impact and the Early Earth
 R. A. F. Grieve and E. M. Parmentier

Lunar and Martian Impact Basins: Exposed Records of Terrestrial Bombardment?
 P. H. Schultz

Large Arcuate Scars: A Geological Legacy of the Earth's Accretionary Past
 J. M. Saul

Monday, April 23
4:00-5:30 p.m.

Isotopic Patterns of the Oldest Rocks: What Do They Tell Us About Earlier Times?

G. Lugmair and J. D. Macdougall, Coordinators

Constraints on the Development of the Early Continental Crust from Isotopic Data
 S. B. Jacobsen

Anorthositic Oceanic Crust in the Archean Earth
 E. Jagoutz, J. B. Dawson, S. Hoernes, B. Spettel, and H. Wänke

Tuesday, April 24
8:30-10:00 a.m.

(Continuation of Previous Session)

The Relation Between the Age of the Subducting Slab and the Recycling of Sediment into the Mantle

D. Abbott and S. Hoffman

Dating Archean Zircon by Ion Microprobe—New Light on an Old Problem

I. S. Williams, P. D. Kinny, L. P. Black, W. Compston, D. O. Froude, and T. R. Ireland

Early Evolution of the Crust-Mantle System

K. C. Condie

Tuesday, April 24
10:15-12:30 am

**Igneous, Sedimentary, and Metamorphic Petrology of the Oldest Rocks;
 The Structure of the Oldest Rocks and Implications for the Oldest Continents**

K. Burke, Coordinator

Chemical Constraints on the Evolution of the Archean Continental Crust

B. L. Weaver and J. Tarney

The Aravalli Sequence of Rajasthan, India: A Precambrian Continental Margin?

J. O. Macdougall, R. Willis, G. W. Lugmair, A. B. Roy, and K. Gopalan

Archean Sedimentary Systems and Crustal Evolution

D. R. Lowe

Earth's Primordial Differentiation and Its After-effects

P. H. Warren

Tuesday, April 24
1:30-3:30 p.m.

(Continuation of Previous Session)

W. Kidd, Coordinator

Crustal Radiogenic Heat Production and the Selective Survival of Ancient Continental Crust

P. Morgan

The 3.5 b.y. old Onverwacht Group: A Remnant of Ancient Oceanic Crust

S. Hoffman

Hydrothermal Alteration in the 3.5 b.y. old Onverwacht Group of South Africa

R. Hart and M. J. de Wit

Submarine Hydrothermal Environments as Sites for the Origin and Evolution of Life

S. Hoffman and J. Baross

The Origin of Continental Crust: Outlines of a General Theory

P. D. Lowman, Jr.

A review of Tectonic Aspects of the Limpopo Belt and Other Archean High-Grade Gneissic Terranes

W. S. L. Kidd

Tuesday, April 24
4:00-6:00 p.m.

Volatiles and Early Earth History, Especially Noble Gases

Mark D. Kurz, Coordinator

A Review of Noble Gas Geochemistry in Relation to Early Earth History
 M. D. Kurz

Constraints on Early Atmosphere from Planetary Accretion Processes
 G. Arrhenius

The Degassing History of the Earth; Noble Gas Studies of Archean Cherts and Zero Age Glassy Submarine Basalts
 R. Hart and L. Hogan

Noble Gases and the Early History of the Earth: Inappropriate Paradigms and Assumptions Inhibit Research and Communication
 G. R. Huss and E. C. Alexander, Jr.

The Composition of the Primitive Atmosphere and the Synthesis of Organic Compounds on the Early Earth
 J. L. Bada and S. L. Miller

Carbon and Nitrogen in the Deep Earth
 D. P. Matthey, L. P. Carr, R. H. Carr, I. P. Wright, and C. T. Pillinger

Wednesday, April 25
8:30 a.m.-12:30 p.m.

The Early History of the Earth: Evidence from Accretion Core Formation, Rock Isotopic Composition, Volatiles, Geochemistry, Petrology and Structure of Old Rocks

Frank Richter, Coordinator

Summarizers:

George Wetherill, *Accretion*

Günter Lugmair and J. D. Macdougall, *Isotopes*

Mark Kurz, *Volatiles*

William Kidd, *Geology of Old Rocks*

Richard Grieve, *Impacts*

Summary of Technical Sessions

(Prepared by K. Burke, R. A. F. Grieve, W. Kidd, and J. D. MacDougall)

Accretion of the Earth: Processes Accompanying and Immediately Following Accretion

Papers presented in this session indicated that a great deal happened in the first 10^4 years of Earth history. This interval apparently sufficed not only for accretion but also for formation of core, mantle, and a crust as well as atmosphere and ocean, these last two perhaps being added to the Earth late in the first 10^4 years of its history. The following questions point out the problems to which attention was drawn in discussion.

1. **What was the size distribution in the population of planetesimals from which the Earth was formed?** *Safronov's* (1969) suggestion that the largest object exceeded the size of the next by a factor of 1000 seems perhaps less probable than a distribution in which the largest object was ten times larger than the next.

2. **Are objects the size of Ceres ($D = 1020$ km) immune from rupture?** It seemed likely that they are gravitationally stable and unlikely to be ruptured except where local tidal forces are unusually large.

3. **What processes in the outer planets are relevant to the formation of the Earth?** For example, could a late flux of volatile material have reached the Earth from the outer planetary region of the solar system?

4. It became clear that there is a need to maintain awareness of the **contingent effects** of events and processes operating early in Earth history. For example, if meteorite contamination of the Moon is rejected, it becomes difficult to accept a late planetesimal veneer on Earth.

5. The topic of **core formation** was addressed from various standpoints. The processes by which siderophile elements are extracted into the core is still unresolved. *Brett* (1984) suggested that the material that is to enter the core might have effectively equilibrated with mantle at shallow depths and consequently low pressures. *Drake* suggested that there was a continuing need for experimental determination of distribution coefficients between metallic and silicate phases.

Stevenson drew attention to the continuing efforts to identify the 10% of the core that is not iron. There is no necessity that this material is siderophile; both H_2 and Si have been suggested as possible components. If it does not turn out that low pressure processes dominate (as *Brett* suggested), then it will take a long time before we can be confident about what the non-ferrous core material might be. There are many reasons why knowledge about the nature of this material is critical; for example, giant gaseous protoplanets should contain a distinctive non-ferrous core component, and establishment of core geochemistry would help in testing whether this mechanism of planet formation operated. Core geochemistry is unlikely to be simple, and our ideas about it have gone through stages of evolution.

Brett, R. (1984) Siderophile elements may not be enriched in the upper mantle: Speculations on protocore separation (abstract), in Lunar and Planetary Science XV, p. 92-93, Lunar and Planetary Institute, Houston.

Safronov, V. S. (1969) Evolution of the protoplanetary cloud and formation of the Earth and the planets, Moscow, Nauka; translated for NASA and NSF by the Israel Program for Scientific Translation, NASA TTF-677 (published in 1972), 206 pp.

Isotope Patterns

The session on isotopic studies produced some lively debate and focused attention both on old problems and new opportunities. Several important themes emerged from the discussions. A major one was the need to define the depleted mantle curve better, particularly in the Archean. This problem is probably best addressed using Nd isotopes; the **maximum** values of Nd isotopic ratios for a given time are crucial parameters since virtually all processes involving crustal material tend to decrease these ratios. The presence of higher than chondritic initial values among Archean crustal rocks implies earlier chemical fractionation in the mantle, presumably the result of crust formation. Questions of the crustal growth rate, the degree of crustal recycling, and the possibility of changes in the recycling regime can potentially all be addressed via the isotope approach, although volume relationships cannot be constrained and at present remain model dependent. The nature of the rocks containing depleted mantle signatures is important for the interpretation of isotopic data; many crustal terranes require a multi-stage mantle-to-crust evolution. This point brings home the artificial nature of the subdivision of Archean problems into "isotopes," or "geology," or other such categories and the danger of approaching the problem from a single viewpoint.

A new viewpoint discussed at the workshop (which nicely illustrated that there are novel ways to look at old problems) was the possibility of using mantle xenoliths as probes of the Archean crust. Oxygen and radiogenic isotopes in eclogites from African kimberlites suggest that these may be relatively undisturbed fragments of subducted Archean oceanic crust. If so, studies of such samples may provide data not only about the isotopic composition of the early mantle, but also about the thermal and/or dynamic history of the subcontinental lithosphere.

An aspect of Archean isotope studies that requires continuing effort is the search for the oldest crustal material. In addition to examining identified old terranes for enclaves of yet older material, it appears that extracting detrital phases, such as zircon, from ancient sediments is a promising approach. Such a technique has apparently uncovered the oldest dated terrestrial material—the important problem now being how to extract information about the parent rocks from individual grains. Careful age determinations, for these and other Archean materials, especially when they are closely tied to field and other geochemical studies, are necessary if the details of early crustal evolution are to be worked out.

Information From Other Planets Relevant to the Early Crustal Evolution of the Earth

As the oldest record of terrestrial geologic processes is preserved in rocks with ages of approximately 3.8 b.y., it is appropriate to consider the geologic record of other planetary bodies that have preserved portions of older crust and to consider the relevance of that record for the earliest geologic history of the Earth. The planetary record indicates that the terrestrial planets were subjected to a period of intense bombardment by residual planetesimal bodies during their early history and that, on bodies such as the Moon and Mercury, impact was a dominant early geologic process.

There is no obvious evidence of this early, high impact flux preserved on Earth. The imagery available for the Moon and Mars serve as references for establishing the number of major impacts that may have occurred on the early Earth after corrections are made for variations in planetary size, gravity, and impact conditions. If the Earth had a crust in the first approximately 0.8 b.y. of its history, it is expected that several tens to perhaps a hundred 1000-km-sized or larger impact basins may have been formed on the early Earth. Evidence from the Moon and Mars indicates that basin-forming impacts created a sizable surface

topography, crustal compositional inhomogeneities, and relatively long lived thermal anomalies. They also served to localize endogenic geologic activity, leaving a structural imprint that controlled the occurrence of later volcanic, tectonic, and, in the case of Mars, erosional and depositional processes. A major question is whether similar impact-related phenomena were geologically important on the more thermally and tectonically active early Earth. Initial thermal-tectonic modeling of the post impact history of large impact basins on the early Earth suggests that the impact induced thermal anomalies may be relatively long-lived and large amounts of impact-induced volcanism may follow adiabatic decompression from deep uplifts to basin formation. It is likely, however, that such basins with their associated fill would be dynamically unstable on a body with a thin lithosphere such as the Earth. While basin-sized events may have promoted crustal heterogeneity and locally influenced mantle convection, the degree to which large impacts influenced craton stabilization and crustal evolution compared to normal endogenic process on the early Earth is still problematical and requires additional modeling.

Observational data indicate that Mars has had a long and complex geologic history and that a variety of endogenic and exogenic geologic processes have shaped its surface. It is geologically intermediate between bodies such as the Moon, where impact dominated early crustal and surface evolution, and bodies such as the highly active Earth, where the recorded history is dominated by internal processes. As an early Earth analog, however, Mars fails in that it has a continuous lithospheric shell. Venus, which is similar in size to the Earth, may provide constraints on potential volcanic-tectonic styles for the early Earth, as well as an example of the relative importance and the interaction of endogenic and exogenic geologic process. Currently available radar imagery, however, is insufficient to adequately evaluate Venus as an early Earth analog.

Geology of Old Rocks

Contributions in this section cover a varied, but by no means comprehensive set of topics bearing on Archean geology, as did the discussions at the workshop. The geology of the old rocks, unlike the isotopic evidence, reflects the Earth's behavior at the time they were formed or deformed. The rocks have less to say about the time between their formation and the time of accretion than the isotopic systems and external constraints. While the geology provides a goal for pre-old rock modelers to aim for, it seems that the constraints it, by itself, places on the subject of the title of this workshop are somewhat weak, except for ideas that suggest the lithosphere was absent (magma ocean) or behaving very differently compared with younger times (e.g., dominated by impact and consequent processes) when the admittedly very limited and battered sample of the older Archean was formed.

Of particular interest in this section of the workshop is the conclusion of Morgan that the sample of Archean crust preserved for our inspection is seriously skewed. Crust having a low content of heat-producing elements (i.e., K, U, Th) when formed has been preferentially preserved, while Archean crust with higher contents of heat-producing elements is lacking and, it is inferred, has been preferentially reworked. This is thought to have occurred because a significantly thinner lithosphere was associated with the steeper thermal gradient forced by the hotter crust, which was therefore more susceptible to attack by tectonic processes (e.g., rifting and subsequent involvement in convergent processes) than other areas with thicker lithosphere. Given the small overall area of Archean crust that is preserved in subsequently unmodified form, this result suggests that it may be inappropriate to read much significance into the present proportions and age distribution of the greenstone granodiorite versus the high grade gneissic Archean terranes.

A general question that arose from the presentations was whether the oldest Archean rocks, including the sediments (Isua and similar sequences), are essentially similar to younger Archean sequences, or

whether there are significant differences between older and younger suites. If there are real and significant differences, are they any more than might be expected to arise from continental growth generated by subduction and arc collision through that time? It may be difficult to establish unambiguous answers to the question, given the small number of the older belts and the likelihood that a skewed sample has been preserved (for the reasons presented at this workshop by Morgan). However, Dymek's geochemical identification of Superior-type iron-formation in the Isua sequence is a case where a previously unsuspected resemblance to younger rocks has been found.

Structures preserved in the oldest rocks are similar in type and large-scale arrangement to younger rocks last deformed and metamorphosed in zones of continental collision involving substantial crustal thickening. As such, the structures reflect significant lateral motions of the lithosphere, which, as McKenzie and others pointed out many years ago, are the surface expression of large-scale convection in the mantle. Whether there was a time in the early history of the Earth when such lateral motion did not take place, and heat was removed largely by cooling large lava lakes, as Stevenson suggests was possible (q.v. Accretion section), it is an interesting question; the geology implies that by 3.8 b.y., lateral motions and subduction were in operation. In particular, the production of tonalitic magmas (see Weaver and Tarney's presentation) requires the recycling of hydrated basalts into the mantle, most easily accomplished by subduction of oceanic lithosphere.

Of the similarities raised between aspects of Archean and younger terranes, another interesting and unexpected result is the repeatable high-temperature seawater $\delta^{18}\text{O}$ values obtained from the Onverwacht komatiites by Hoffman. Also surprising was the fairly general ignorance, revealed by the discussion of these results, of the totally altered, hydrated nature of the komatiites of this and other greenstone belts. An intriguing suggestion was made by Warren to account for the ultramafic komatiites. Very efficient early outgassing is proposed to have produced a dry upper mantle that only became generally hydrated by the end of the Archean. Such dry mantle would melt at appropriately higher temperatures and would avoid some difficulties of earlier models. Whether the idea is testable, given the generally altered nature of the komatiites, is more difficult to judge. The identification of a particular tectonic setting for the eruption of the ultramafic komatiites would enable better constraints to be placed on their significance. It seems most likely that this setting is oceanic, but it is presently difficult to identify it with confidence any more precisely. In the Barberton belt, through the work of DeWit, a fracture zone setting looks plausible; it would be useful to confirm this for other examples. The open question of their relationship to the layered anorthosite suites, if both are versions of Archean oceanic crust, is a reminder of our current lack of overall understanding. In discussion, mention was made of the report some years ago that a thick section of komatiites overlies a thin section of shallow water sediments resting unconformably on tonalitic gneissic basement at the eastern margin of the Bulungwe greenstone belt of Zimbabwe. There is a strong possibility that a major thrust separates the komatiites from the sediments, so it should not be assumed from this example that komatiites were erupted onto continental crust.

Most of the presentations in this section favored or at least were not evidence against the operation of plate tectonic processes (or processes very similar to plate tectonics) by the time the older rock record began accumulating. This is in striking contrast to most views expressed in similar meetings only a few years ago, although it would not be fair to say that everyone agreed on this topic. Hargraves pointed out in discussion that it may be dangerous to infer too much similarity between Phanerozoic and Archean tectonics, in that differences will only be found if they are sought out. However, large-scale lateral motions of the Archean lithosphere are required to explain the dominant early nappe and related ductile structures seen in both gneissic and greenstone terranes. Resemblances between the first-order tectonic and structural aspects of these terranes and those known to have been created by plate tectonics in the Proterozoic and Phanerozoic are much greater than any differences that have been emphasized in the past. Others therefore prefer to test the hypothesis that plate tectonic processes were operating in the Archean, rather than to propose some other system that is not operating now and that was capable of simulating

the effects of plate tectonics to a remarkable degree. Clearly, there are differences in the degree of development of particular features and in proportions of various lithic assemblages between Archean and younger orogenic belts; there are also a few items that are unique or are strongly partitioned to Archean (ultramafic komatiites, layered calcic anorthosites) or post-Archean time (alkalic magmatism). Investigation of these differences will continue to be of fundamental importance in helping to understand the Archean Earth. As far as the preserved rock record goes, however, it does not seem necessary to invoke large-scale, first-order differences in tectonics between Archean and younger times.

ABSTRACTS

THE RELATION BETWEEN THE AGE OF THE SUBDUCTING SLAB
AND THE RECYCLING OF SEDIMENTS INTO THE MANTLE

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The recycling of sediments into the mantle has become an important issue because recent papers have suggested that the geochemical inverse models of the evolution of radiogenic isotope abundances over the history of the Earth have nonunique solutions (1,2). Both the recycling of continent-derived sediments into the mantle and mixing in the mantle could produce similar geochemical effects in the mean isotopic ratios of new igneous material emplaced in continents. Recent models of Archaean heat flow (3) and of plate tectonics during early Earth history (4) have demonstrated that higher internal heat production of the early Earth was mainly dissipated through a higher creation rate of oceanic lithosphere. If the seafloor creation rate was higher on the early Earth, then the residence time of any one piece of oceanic lithosphere on the surface would have been shorter. It is possible that a higher rate of recycling of oceanic lithosphere into the mantle could have resulted in some transport of sediment into the mantle. Such transport would be dependent upon the rate of reheating of the upper surface of the slab in the subduction zone, and this in turn is dependent upon the age of the oceanic lithosphere at subduction (Figs. 1,2). Because the subducting slab can act as a conveyor belt transporting sedimentary material into the mantle, we are concerned with the rate at which the upper ten kilometers of the subducting plate attains temperatures above the melting points of sedimentary materials.

Mathematical Model: We use a conductive model for the reheating of a slab with an initially linear increase of temperature with depth as the slab enters a medium of constant potential temperature (Fig. 1; Ref. 5,6). Estimates of the isotherms in the slab at the lithosphere-asthenosphere boundary are derived by running the model for conditions in the asthenosphere while assuming that the rate of heating in the lithosphere is 1.7 times lower than in the asthenosphere (7). This produces only approximate results, but they are sufficiently accurate for a rough estimate of the effect of age of the subducting slab on the probability of subducting sediments into the underlying asthenosphere.

Modelling Results: The results show that the temperature gradient of the subducting slab at the asthenosphere-lithosphere boundary is greatest for the youngest age of subducting lithosphere and for the case where oceanic lithosphere subducts beneath continental lithosphere (Fig. 2). Because the youngest lithosphere is also likely to have the thinnest sediment cover, the probability of melting or underplating all of the subducted sediments is greatest for the youngest slabs (Fig. 3). Therefore, significant quantities of sediment are only likely to be returned to the mantle if the subducting slab is relatively old. The more siliceous sediments, such as arkoses, clean sandstones, dacitic to rhyolitic volcanoclastics, and siliceous carbonates, should melt completely at about 900°C. Therefore, siliceous sediments should not be transported into the mantle until the maximum age of subducting slabs is greater than ~60 m.y. in the case of ocean-ocean subduction and ~76 m.y. in the case of ocean-continent subduction. The estimated liquidus for more basic sediments, such as red clays and andesitic and basaltic volcanoclastics, is ~1140°C. Therefore, more basic sediments could begin to be recycled into the mantle when the maximum age of the subducting slab is greater than ~27 m.y.

for O-O subduction and ~ 40 m.y. for O-C subduction. Those sediments with the greatest concentrations of volatile and incompatible elements, such as Sr and Rb, tend to be the most siliceous sediments with the lowest melting points. Consequently, Sr isotope ratios began to be affected by the recycling of sediments into the mantle when the maximum age of the subducting lithosphere approached 60 to 80 m.y. According to our model of continental accretion (4), this occurred roughly 2.6 to 3.0 b.y. ago. This is very close to the time of a major change in the slope of the Sr isotope evolution curve (8,9).

Implications for Early Earth History and Geochemical Models: If, as we propose (4), the maximum age of the subducting oceanic lithosphere was less than 60 m.y. prior to ~ 3 b.y. ago, then all siliceous sediments which were subducted melted before reaching the asthenosphere. The nature of Archaean oceanic sedimentation is not well-known, but some conclusions can be drawn on the basis of our model (4). It is apparent that the relative abundance of mature terrigenous clastics is dependent upon the size of continents and the nature of continental weathering. In the earliest Archaean, continental masses were probably small. It is therefore probable that Archaean oceanic sedimentation was dominated by island arc sources and that most subducted sediment consisted of immature clastics of calc-alkaline composition. Melting of these sediments in subduction zones would explain the origin of potassic granites which have REE patterns indicating that they were derived from granodioritic sources.

Furthermore, petrological evidence cited in Ref. 4 indicates that the oceanic slab itself would produce small partial melts at relatively shallow depths and temperatures less than $\sim 800^\circ\text{C}$. There is the potential for mixing of magmas derived from the slab and those derived from subducted sediments which could produce complicated and confusing REE patterns in the resulting igneous rocks. However, the calc-alkaline suite of the earliest Archaean should be the least "contaminated" by recycled sediment and is the most likely to have been totally derived from the subducting slab itself. We are proposing that, as the average age of subducting lithosphere increased over the history of the Earth (4), sedimentary contributions to arc magmatism, continental growth, and mantle contamination increased.

In terms of geochemical models of continental accretion, it appears that some of the problems of nonuniqueness resulting from sediment recycling, magma mixing, and contamination of mantle reservoirs may not apply to data from the earliest Archaean. It may therefore be possible to model continental accretion using isotopic data in conjunction with an evolutionary plate tectonic model rather than a poorly-constrained inverse model.

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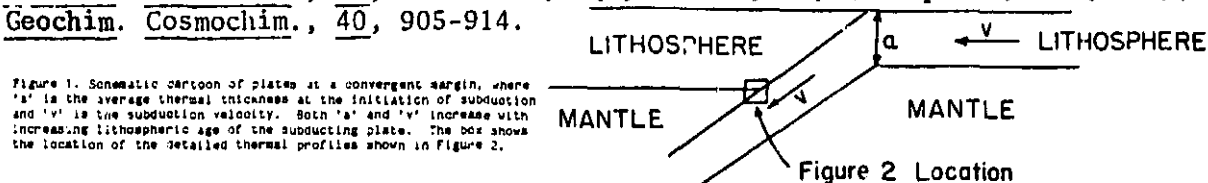


Figure 1. Schematic cartoon of plates at a convergent margin, where 'a' is the average thermal thickness at the initiation of subduction and 'v' is the subduction velocity. Both 'a' and 'v' increase with increasing lithospheric age of the subducting plate. The box shows the location of the detailed thermal profiles shown in Figure 2.

Figure 2 Location

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Figure 2. Thermal profiles for the upper 10 km of subducting oceanic lithosphere calculated using the quantities 'a' and 'v' from Figure 1. (top) Ocean-Continent subduction where oceanic lithosphere subducts beneath continental lithosphere with a thickness of 125 km. Profiles were calculated for subducting oceanic plates of the following ages: 10, 20, 30, 40, 50, 60, 80 and 100 m.y. when the slab enters the trench. (bottom) Ocean-Ocean subduction where oceanic lithosphere subducts beneath oceanic lithosphere of the same age and thermal thickness. Profiles were calculated for ages of 10, 20, 30, 40, 60, and 100 m.y.

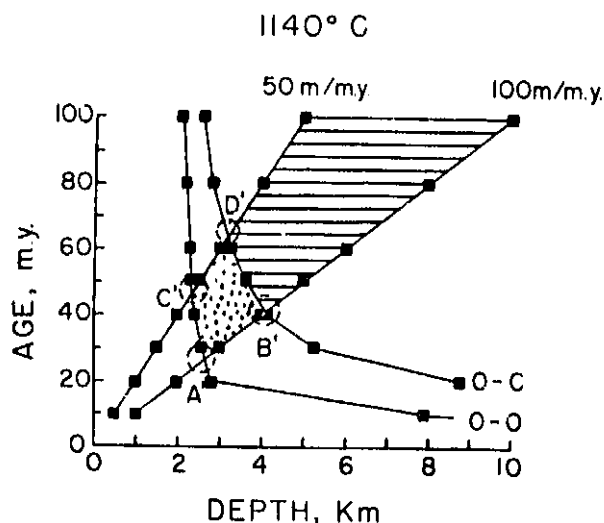
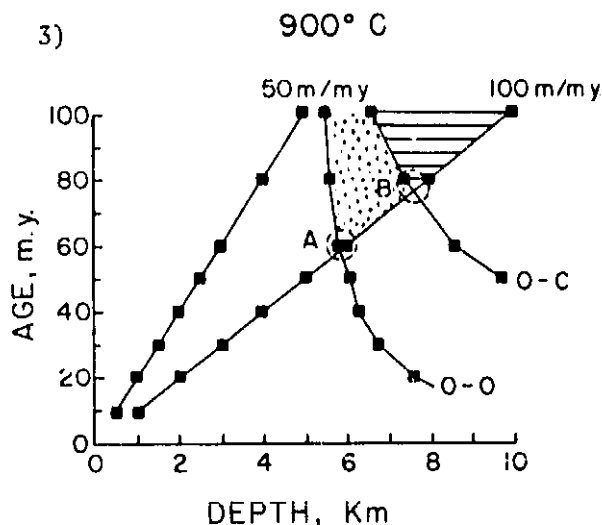
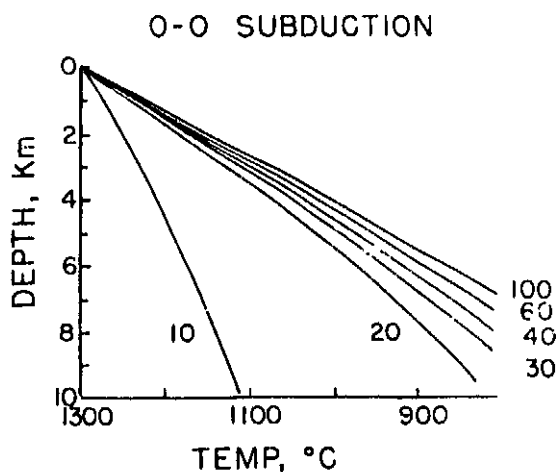
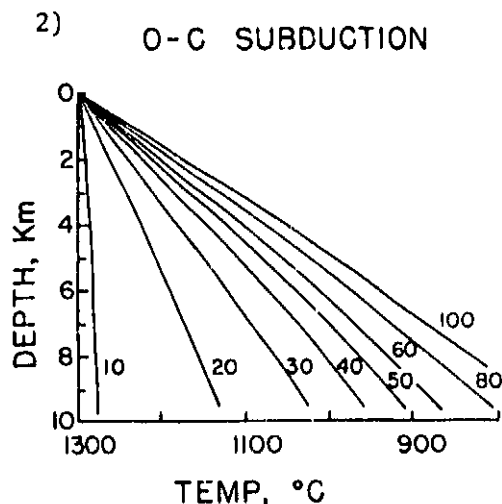


Figure 3. Relationships between depths to isotherms in the subducting slab, sedimentation rate, and the probability that sediments will be subducted into the mantle. Isotherms chosen, 900° and 1140° C, represent the approximate hydrous liquid of quartz-rich turbidites and red clays respectively. (top) Depths of the 900° C isotherm below the top of the subducting slab as a function of lithospheric age for ocean-ocean (O-O) and ocean-continent (O-C) subduction. Points A and B show the intersections of the isotherms with the base of a sediment pile resulting from an average sedimentation rate of 100m/m.y. The dotted and lined regions represent those circumstances where some sedimentary material will be subducted into the mantle. (bottom) Depths of the 1140° C isotherm for the same conditions as top figure. Points A', B', C', D' show the intersections of the isotherm with the base of sediment piles resulting from sedimentation rates of both 50m/m.y. and 100m/m.y.

CONSTRAINTS ON EARLY ATMOSPHERE FROM PLANETARY ACCRETION PROCESSES. G. Arrhenius, Scripps Inst. Oceanogr., La Jolla, CA 92093.

The limited insights in accretion processes that existed in the pre-Apollo era permitted a wide range of assumptions concerning the composition and evolution of the early atmosphere. Most concepts were based on the belief that present day volcanism represents late outgassing of volatiles incorporated in the Earth at its formation. The formation process was somewhat vaguely visualized as a quiet geocentric settling of dust in a giant gas cloud of solar composition to form a primitive planet consisting of an undifferentiated mixture of volatile and refractory materials. Radiogenic heating would then gradually force a release of primordial volatiles from the interior of the planet.

Evidence for composition and dynamics of release of the earliest volatiles was in lack of anything better being sought from the most ancient sediments known at the time, although they were recognized to be younger than three billion years. Origin of life on Earth was considered to require a lasting atmosphere with hydrogen and methane as major components.

The new observations together with theoretical studies generated by the space program and by systematic exploration of the Earth's oceanic crust changed the climate of opinion in which the questions of the primordial atmosphere are discussed. Even though existing evidence does not permit conclusive choice of any specific scenario, the acceptance of specific model elements now forces the consideration of a series of consequences, some of which may be tested by observation.

One of the realizations from lunar exploration is the effective degassing of material falling on planetary bodies with escape velocities exceeding one or a few km/sec. If accretion of the Earth, as is generally believed, took place from impacting planetesimals, the growth of the planet should have been accompanied by the concurrent separation, growth and increasing gravitational retention of an atmosphere consisting of the volatile components released at the conversion of kinetic energy into heat. Furthermore all current models for planetesimal growth imply that the major mass falls as large bodies which penetrate even the late stage atmosphere and explode at impact. Rather than assuming as in the past that a large reservoir of primordial volatiles became buried in the Earth's interior and would be continually emerging, the new inquiry concerns the ploughing down and recycling over geological time by crustal subduction of volatiles, effectively concentrated in a primitive atmosphere-hydrosphere at accretion. This process includes the gradual and irreversible loss of reactive volatiles such as oxygen from the ocean-atmosphere system into the mantle of the Earth. The question arises if any "juvenile" emerging components exist at all in such quantities that they can be traced and suggests a re-evaluation of the sources of helium-3 emanating from the crust. Excess helium-3 is now generally assumed to be primordial and thus to derive from a mantle source never previously degassed. It is difficult to reconcile what is now known of impact degassing on the Moon and planets with a mechanism that would allow

Arrhenius, G.

the most diffusive of all non-reactive elements to be selectively left behind in the solid residue from the surface explosions at planetary accretion.

Two alternatives would seem possible. One would indicate the inner core of the Earth as a source of primordial helium-3. Using the estimated present day flux at the surface of the planet, it can be shown that the total amount escaped over the life of the Earth if distributed over the volume of the inner core would give a concentration of helium-3 compatible with the average for chondritic meteorites. When it had grown to the size of the present inner core, the embryonic Earth had sufficiently low escape velocity ($<1 \text{ km sec}^{-1}$) to permit effective burial of volatiles. Particularly elements with rapid molecular diffusion such as helium could gradually and selectively escape into the liquid outer core and by convection be brought to the upper mantle and crust.

Another alternative, which could be verified experimentally, would imply that there exists a helium-3 source in the oceanic crust with a strength such as to give the observed flux in active centers. An observation of possible importance in this context concerns the extraction of lithium and boron from sea water in the thermally forced circulation system in the oceanic crust, where these elements reach highly localized concentrations exceeding 100 ppm in veins of saponite clay (Andrews, 1978). Uranium, providing a source for neutrons (from spontaneous fission of ^{238}U) and for alpha particles, is similarly concentrated into this formation. The crucial feature is the strong concentration of the reactive nuclides together in microscopic domains, comparable in size with the range of the nuclear decay particles.

It is questionable whether this constellation alone can account for the observed flux of ^3He , given the currently accepted capture cross-sections for the nuclear reactions leading to the formation of tritium and helium-3. These values may, however, be too low (Lal, 1984; pers. comm.). If the crustal alternative for the origin of helium-3 were borne out, no indication would remain of a flux of primordial volatiles from the mantle, in agreement with the suggestions from the Moon and from planetary growth dynamics. Methane and other hydrocarbons could have a deep earth origin in so far as they may derive from reaction of metal carbides or graphite with water but are not primordial gases by virtue of such an origin.

The picture that we visualize is thus a primordial atmosphere and ocean, co-eval with and separated from the growing planet and with a composition similar to the volatile fraction of ordinary chondrites. Over geological time the originally volatile-depleted lunar-type mantle gradually becomes contaminated by hydration and subduction of crust which has been in contact with the ocean-atmosphere. The mantle, instead of being a source of primordial volatiles thus acts as a sink, gradually depleting the hydrosphere by oxidation of iron with residual water in subducted crust. This reaction is driven by the continuous escape of hydrogen into space - the planet is in this respect an open system.

The rate of power dissipation at the growth of the planet is sensitive to the content at any given time, of source material in the Earth's region of the solar system and thus to the rate and the termination of influx of interstellar matter into the critical velocity region of the terrestrial planets. In one scenario the runaway acceleration of growth exhausts the Earth's source jet streams when the Earth has reached the size of the core; the continued accretion growth of the mantle then remains choked at a low rate. Eventually the ending supply of interstellar source material in the terrestrial region leads to the termination of growth of the planet after a period of accretion of the order of 10^6 years.

In another model, representing the extreme, all of the Earth's source material is present in the terrestrial region of interplanetary space at the outset. Runaway accretion then accelerates catastrophically and culminates at the end of planet growth, leaving the Earth at a temperature exceeding 10,000 K and insulated against rapid heat loss by a dense atmosphere containing practically the entire volatile inventory. Effective gravitational separation of metal phases into the core would be expected. However, the fact that metallic alloys, such as nickel-iron, remain in the lower crust and upper mantle seems to indicate that the planet has never been at one time completely melted. This supports the concept of accretion proceeding by the hot spot mechanism, and with an ocean beginning to form together with a retained atmosphere somewhere between Moon- and Mars-size of the embryonic Earth.

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THE COMPOSITION OF THE PRIMITIVE ATMOSPHERE AND THE SYNTHESIS
OF ORGANIC COMPOUNDS ON THE EARLY EARTH

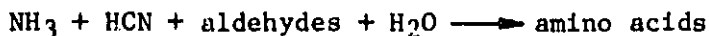
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The generally accepted theory for the origin of life on the earth requires that a large variety of organic compounds be present to form the first living organisms and to provide the energy sources for primitive life either directly or through various fermentation reactions. This can provide a strong constraint on discussions of the formation of the earth and on the composition of the primitive atmosphere. This factor has not been seriously taken into account in many recent papers dealing with the earth's primitive atmosphere [for example, see ref. (1)]. Our objective here is to emphasize that in order for substantial amounts of organic compounds to have been present on the prebiological earth, certain conditions must have existed.

There is a large body of literature on the prebiotic synthesis of organic compounds in various postulated atmospheres [for a review, see ref. (2)]. In this mixture of abiotically synthesized organic compounds, the amino acids are of special interest since they are utilized by modern organisms to synthesize structural materials and a large array of catalytic peptides. We will thus tend to focus our attention on these compounds.

The first abiotic experiments that gave high yields of amino acids used strongly reducing conditions and an electric discharge as an energy source (2). Abelson (3) subsequently used various mixtures of CO and CO₂ with H₂, N₂, N₂O, and found amino acids were synthesized only when the system had an excess of H₂. More recent experiments have studied these gas mixtures systematically (4,5).

It is apparent that the most effective abiotic synthetic pathway on the early earth would involve an atmosphere which contains reduced species such as methane, ammonia and hydrogen. The reaction responsible for amino acid synthesis on the early earth if reducing conditions dominated is shown below (2):



This synthesis (Strecker synthesis) would have taken place in an aqueous environment (oceans) but the precursors, hydrogen cyanide and aldehydes, would have been produced in the atmosphere by the action of various energy sources such as spark discharges on an atmosphere containing reduced gases.

The yields of amino acids produced by sparking various mixtures of CH₄, CO, CO₂, H₂, H₂O, N₂ and NH₃, have been studied in detail (4,5). The CH₄ mixtures gave a rich variety of amino acids while with CO and CO₂ essentially only glycine was synthesized. Very low yields of amino acids were obtained when the H₂/CO₂ ratio was zero, whereas high yields were obtained at high H₂/CO and H₂/CO₂ ratios.

The prebiotic synthesis of organic compounds has never been accomplished in the presence of molecular O₂. Photosynthesis by blue-green algae and

Bada, J. L.

plants, of course, do such syntheses but this is a highly evolved biochemical system, the precursor of which was the anaerobic photosynthesis by bacteria (2). An additional problem with the presence of O_2 in the early earth's atmosphere is that O_2 will decompose prebiotic organic compounds. Estimates of the rate of decomposition (6) indicate that most amino acids are stable in the absence of oxygen for periods as long as 10^8 years at temperature in the range 0° to 25° . At higher temperatures, amino acids are less stable. Also amino acids are rapidly decomposed when oxygen is present. The estimated half-life for the oxidative decomposition of alanine at $0^\circ C$ is about 10^6 years, which is 10^8 times faster than the rate under anoxic conditions (7). Moreover, metal ions are very effective catalysts for the oxidation of amino acids. The estimated half-life for the metal ion catalyzed oxidation of alanine in the modern ocean is about 350 years (7). Thus, it is apparent that in order for amino acids and probably other organic compounds to accumulate, an anoxic early earth would have been required unless life arose very rapidly.

Because high yields of amino acids are produced using simulated primitive atmospheres containing reduced components, we feel that this is a good indication of the type of atmosphere needed for amino acid synthesis on the early earth. This conclusion implies that during at least some period of the early earth's history reduced gases such as CH_4 , NH_3 and H_2 must have been present in the atmosphere. The main arguments for the absence or low abundance of reduced gases in the primitive earth's atmosphere are that CH_4 and NH_3 are photochemically unstable, and that a high concentration of an effective greenhouse molecule such as CO_2 is required in order to keep the early earth's surface temperature above the freezing point of water (1). However, CH_4 and NH_3 are both greenhouse gases (8) and their presence in the early earth's atmosphere would have thus helped maintain a mean surface temperature greater than $0^\circ C$ (9). Moreover, the photochemical decomposition of CH_4 and NH_3 could have been significantly retarded if the primitive earth's atmosphere contained particulates and haze (10).

For abiotic organic compound synthesis to have occurred on the early earth, high concentrations of the required reduced gases in the atmosphere are not necessary. Methane pressures of 10^{-5} and 10^{-2} atm would have been sufficient. Most of the nitrogen in the atmosphere would be N_2 , but small amounts (10^{-6} to 10^{-8} atm) of NH_3 are required for the synthesis of amino acids (11,12).

Preservation of organic compounds on the early earth so they could accumulate to levels where they were sufficiently abundant for eventual utilization by the first primitive organisms would have been possible only if the oxygen content of the atmosphere was low, and surface temperatures on the earth were relatively cool.

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CONDITIONS DURING FORMATION OF THE EARTH. A.G.W. Cameron, Harvard-Smithsonian Center for Astrophysics.

I have made a number of studies of the formation of the primitive solar nebula and its early evolution which have a bearing on the conditions under which the accumulation of the Earth took place. None of these studies were primarily directed toward explaining any purely terrestrial phenomenon, so there may be some interest in examining their predictions for the formation of the Earth. Some of these predictions depart very much from what one might expect from a backwards extrapolation of currently active processes.

(1) Time scale for formation of the sun. In dense molecular clouds in the interstellar medium where star formation is taking place, the newly-formed young emission-line stars (T Tauri stars) are distributed in a well-defined fashion in an H-R diagram (in which the luminosity of the stars is plotted against their surface temperatures). The evolution of these stars causes them to move at first primarily in a direction from high to low luminosity with little change in surface temperature. Recent studies (1,2) have clarified the conditions required to form the T Tauri stars in their initial positions in the H-R diagram. For a star of about one solar mass, the rate of accretion near the terminal stage of star formation must be greater than 10^{-5} solar masses per year by at least a factor of several, so that the total accumulation time should not exceed about 10^5 years. If these conditions are not met, not enough energy is stored in the outer layers of the star, and the post-accretion star is not as luminous as it should be.

(2) Turbulence in the infalling gas. Star formation predominantly takes place in dense molecular clouds starting from locally denser regions called cores (3). The initial gas temperature is 10 K and there is also an internal velocity field in excess of thermal which appears random and has typical velocities of about Mach 0.7. This internal velocity field should establish a random total angular momentum vector and it should be amplified when the core undergoes gravitational collapse. This means that the infalling gas will contain large internal turbulence, causing it to deviate from straight radial infall and to be deposited with random local velocities on the primitive solar nebula.

(3) Giant gaseous protoplanets. In the very early stages of the gas infall, when it starts collecting to form the primitive solar nebula, the turbulent velocities will prevent any strong central concentration of the gas, but rather will produce a chaotic and lumpy distribution of gas roughly concentrated to a plane. The initial temperature is expected to be very low. Under these circumstances, many of the lumps can be expected to become gravitationally unstable against forming local structures in hydrostatic equilibrium: giant gaseous protoplanets. With very little mass present, the gas is very much spread out; the radii of the protoplanets exceeds an astronomical unit (4), but after an initial period of sorting themselves out, they are not too likely to collide with one another. Their masses are likely to be about that of Jupiter with a spread of a factor of several. Before too long, the continuing gas infall causes the local temperature to start rising, and the response of the protoplanets in the inner solar system is to have their envelopes thermally evaporated away on quite a short time scale (5). If the interstellar grains imbedded in the gas are able to clump together, which seems likely, then the settling of the grain clumps through the gas will probably take place faster than the envelope evaporation. A Jupiter mass of gas of solar composition will yield one Earth mass of the more refractory elements. Only such elements will be present in the clumps at the temperatures in the protoplanet interiors at the time of envelope evaporation (several hundred degrees (5)).

(4) Protoplanet remnants. As refractory grain clumps settle through the protoplanet gas toward the center, a large amount of gravitational potential energy is released. The total formation time is only several hundred or at most a few thousand years, so that the rocky remnants that are formed will be molten and much of the ambient atmosphere may be rock decomposition products. These rocky remnants will be vigorously thermally convective, but that is unlikely to be able to prevent the denser iron and its dissolved elements from promptly settling out to form a central core. The very high surface temperature of the remnants is likely to maintain the lower layer of the gas in contact with it at or close to the hydrogen escape velocity, so that a vigorous convection zone is likely to keep these surface gases well mixed with the solar

nebula and to sweep away fine dust. I believe that one of these core-mantle remnants is a good candidate to form the nucleus of the Earth, with perhaps 70 or 80 percent of the final Earth mass.

(5) Solar nebula dissipation. Much progress has been made in recent years in understanding the dissipation of accretion disks. If the physics of the viscosity can be specified, then the formalism of Lynden-Bell and Pringle (6) can be used to compute the rate at which the dissipation occurs and the corresponding rates of mass, angular momentum, and energy transport and local energy deposition in the disk. The local energy deposition requires transport in order to allow the energy to flow to the surface and to be radiated away, and so the disk is likely to develop superadiabatic gradients and to be thermally convective. A new theory which relates this convection to the viscosity of the disk has recently been developed by Canuto and his colleagues (7). I have used this physics to construct models of the inner parts of the primitive solar nebula subject to the assumption that the mass flow has become steady there. These models have a range in internal entropy and in the corresponding masses, temperatures and other thermodynamic properties, and in the steady mass flow rates.

(6) The rapid dissipation phase. Solar nebula models with one solar mass at the center and a mass flow rate required to be greater than 10^{-5} solar masses per year have a very interesting property. The temperature exceeds that needed for total evaporation of refractory materials in the disk out to radii well in excess of one astronomical unit. Hence all finely-divided condensed bodies in the inner solar system will be vaporized. Protoplanet remnants will have their surface layers vaporized, and some sort of thermal escape will be established at the tops of the atmospheres so formed. For the more massive bodies the rate of escape will be so small that for all intents and purposes the bodies are stable in the high temperature field. This may not be the case for Mercury, which probably formed with about double its present mass and a normal silicate/iron ratio. I intend to compute the rate of escape of mantle decomposition products as a function of the external temperature for a primitive model of Mercury in order to check this prediction of the model. The expectation is that the silicate mantle of Mercury will be largely evaporated away from the iron core.

(7) The thermal stabilization phase. As the rate of gas infall to the solar nebula diminishes, the diffusion of energy emitted from the protosun progressively warms the outer part of the solar nebula, eventually shutting down the convection zone there. This process spreads inwards, so that eventually the dissipation of the solar nebula ceases and the disk comes into temperature equilibrium with the energy received from the sun and reradiated away to space. These temperatures are considerably cooler than in the rapid dissipation phase. Since convection has ceased, small solid materials can now settle through the disk to collect at midplane. There they will become gravitationally clumped together into bodies of a few kilometers in size (8), becoming asteroids or comet-like bodies, depending on the local temperature. Orbital perturbations with protoplanet remnants will send many of these into the inner solar system. Small bodies not collected into asteroids will suffer gas drag and spiral into the inner solar nebula, where most of them will be collected by the protoplanet remnants there. Meter-sized bodies travel inwards quite rapidly. These bodies contain the elements of medium volatility that were missing from the protoplanet remnants. This accumulation will add significantly to the mass of the inner protoplanet remnants, making them into the planetary bodies that they are today. The time scales here are probably several times 10^5 years, but may be significantly longer.

(8) Primitive planetary atmospheres. It appears necessary for the early solar wind (T Tauri phase wind) to remove the remaining primitive solar nebula, but the time scale on which this can be done is not known. It is known observationally that very young stars can produce strong and very energetic bipolar outflows, and that even the youngest T Tauri stars that can be observed (after the obscuring material has been removed) have relatively slow spins indicating that a great deal of early angular momentum loss has taken place, presumably through the stellar wind. The inner planets show no traces of atmospheric noble gases derived from the solar nebula; rather, the patterns of their abundances reflect the abundances of noble gases absorbed in meteorites. The picture that emerges is one in which the planets were too hot to retain traces of the primitive solar nebula after it was removed, but gradually built up and retained the atmosphere outgassed

Cameron, A.G.W.

from incoming planetesimals. Even much of this atmosphere may be lost through "atmospheric cratering" by incoming planetesimals (9).

(9) Formation of the Moon. The angular momentum in the Earth-Moon system corresponds to that contained in a tangential collision of the protoearth with a body having the mass of Mars. If such a colliding projectile were a protoplanetary remnant, its rapid formation would have assured the prompt separation of its iron core from its mantle. In the collision itself a large amount of mantle rock will be vaporized; it is the pressure gradient established in such a vapor that is responsible for accelerating some of the material into an orbit around the protoearth. This material is strongly dissipative and will spread out and tidally interact with the protoearth in such a way that much of it collects to form the Moon beyond the Roche lobe of the Earth (9). The disk that is formed in orbit is self-gravitating but subject to continuing shear; the primordial atmosphere of the Earth is caused to escape by this disk and the high temperature of the environment, all except a highly mass-fractionated remnant of the xenon; the resulting fractionated pattern of isotopic abundances of the xenon characterizes the atmospheric pattern of xenon today. The collision adds about ten percent to the mass of the protoearth and deposits about 3×10^{38} ergs of new thermal energy in the Earth, remelting any parts of the mantle or crust that had solidified. The present terrestrial atmosphere is then the atmosphere acquired through late accretion following the formation of the Moon. All of this is expected to happen for dynamical reasons at an Earth age of roughly 10^8 years for a projectile that is in a reasonably eccentric Earth-crossing orbit.

(10) Early cooling of the Earth If the very hot Earth had a vigorously convecting mantle and the surface rocks were also molten, with the surface radiation free to escape into space, then the Earth would only take a few thousand years to undergo a major cooling including a great deal of mantle solidification. Things are not so simple, since the surface rocks cannot radiate directly to space; their radiation will be absorbed by the atmosphere, which initially will consist mostly of steam. High in the atmosphere clouds will form, and the temperature at the cloud tops is a better measure of the effective temperature with which the Earth radiates to space. The Earth thus cools relatively slowly, and mantle solidification is likely to be spread over hundreds of millions of years.

(11) Summary. The formation of the Earth starts with gravitational instabilities in the gas of the primitive solar nebula, which form giant gaseous protoplanets. The envelopes of these are thermally evaporated, but solids can gravitationally settle to the center of the protoplanets, and most of the core and mantle of the Earth were formed at that time (age about 10^3 years). The protoearth then survived a period of high temperatures in the surrounding gas (age about 10^5 years). While and after the gas was thermally stabilized and removed from the solar system, the Earth grew by bombardment of planetesimals and its early atmosphere was established. A late major collision led to formation of the Moon and loss of the primordial atmosphere (age about 10^8 years). The present atmosphere resulted from still later accretion. The presence of this atmosphere spreads out the mantle cooling and solidification over hundreds of millions of years.

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EARLY EVOLUTION OF THE CRUST-MANTLE SYSTEM

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Nd isotopic data indicate that most Archean igneous rocks including compositions ranging from komatiite to tonalite are derived from undepleted or depleted upper mantle sources (1). If sampling is representative, only a few require enriched sources. A major unresolved question is the fate of the material removed from the upper mantle leaving early depleted sources as residue (2). One possibility is that widespread depletion of the early mantle resulted from a period of early degassing and magmatism. Rare gas isotopic data, in particular $^{129}\text{Xe}/^{130}\text{Xe}$ ratios, seem to demand that the upper mantle was extensively degassed at or before 4.4 b.y. and this led to rapid growth of the atmosphere and oceans (3). The lower mantle, however, was not significantly degassed during this event. It is likely that such widespread degassing and magmatism of the upper mantle transferred significant quantities of incompatible elements into the uppermost mantle or crust. Once formed, such an enriched fraction should resist recycling into the mantle and collect at or near the earth's surface. One possibility is that it collects chiefly in a zone of partial melting, analogous to the present low-velocity zone at the base of the lithosphere.

Most of the early enriched mantle may have been recycled and mixed with depleted mantle and such mixing may explain the absence of modern basalts derived from either highly depleted or highly enriched mantle sources. An alternate explanation for the absence of such sources is that significant quantities of continental sediments have been recycled back into the mantle (4).

Geochemical model studies of Archean tonalites indicate that their sources are enriched mafic sources rather than depleted sources as indicated by most Nd isotopic data (5). It would seem that the only way to accommodate both sets of data is for the enrichment events to be short-lived such that they are not detectable in Nd isotopic ratios (i.e., ≤ 150 m.y.). Such short-lived enrichment events must have begun by 3.8 b.y., the age of the oldest tonalites. It is probable that early tonalitic continents formed over convective downcurrents where the least steep geotherms existed. Nd isotopic data strongly suggest that the continents grew rapidly to 50-80% of their present volume chiefly at 2.7 (± 0.1) b.y. (6) and that they were extracted chiefly or entirely from the upper 50% of the mantle (7).

The following model, which is not specific in terms of tectonic regime, accommodates isotopic, geochemical, and geothermal constraints for the evolution of the early crust-mantle system. Whether or not a magma ocean existed on the primordial earth is not known. If it did, by analogy with the moon (8), it probably crystallized rapidly during the first 100 m.y. after planetary accretion. During this same period of time the upper 50% of mantle was rapidly degassed, producing depleted and enriched zones (DMR and EMR, respectively) (Fig. 1; Table 1). After and perhaps during this time, komatiitic crust and lithosphere are produced from DMR at convective upcurrents. This lithosphere cools rapidly, becomes negatively buoyant in < 20 m.y., and is recycled back into the mantle at sinks over downcurrents. Partial melting of descending komatiite produces depleted basalts (DBAS) (Table 1) which rise and collect over sinks.

As the earth continues to cool, both komatiite and basalt are produced at upcurrents. By 2.8-4.0 b.y., locally thickened basaltic plateaus over sinks

Condie, K. C.

(with depleted Nd isotopic signatures) begin to melt (at ~ 50 km thickness) producing small amounts of tonalite magma which rise buoyantly, intruding near-surface basalts and komatiites. Just prior to melting, these basalts are enriched by metasomatic fluids from a mixed mantle source comprised of DMR and EMR. Mixing occurs in the sinks.

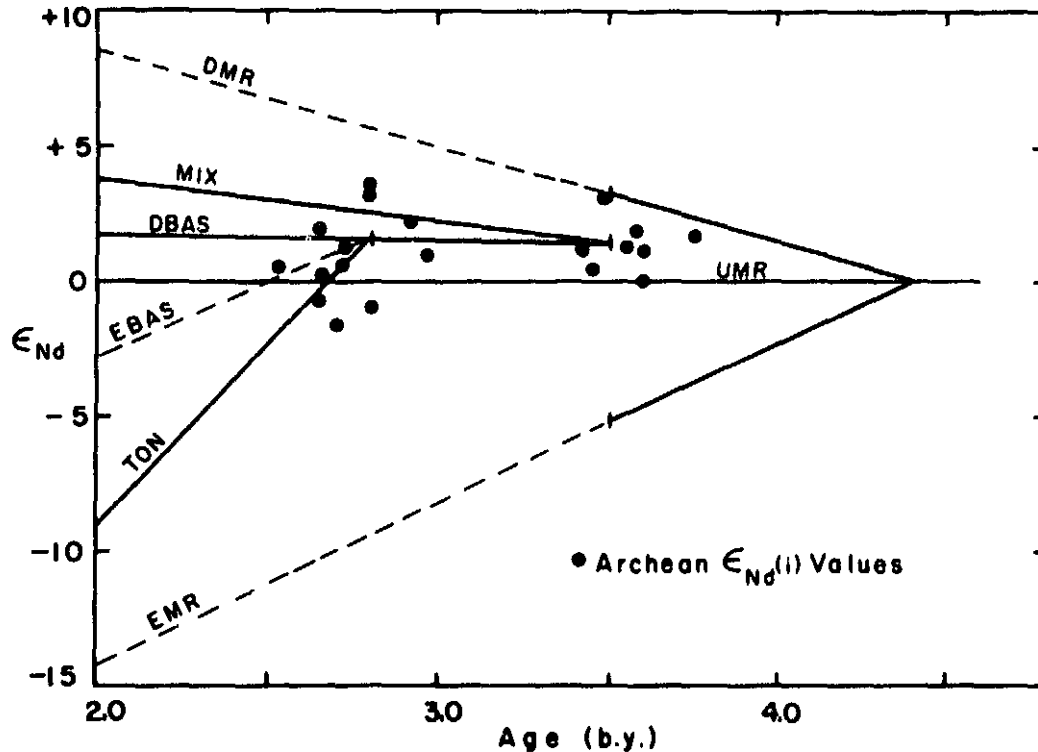


Fig. 1. ϵ_{Nd} growth curves for various mantle and crustal reservoirs. DMR, depleted mantle reservoir; EMR, enriched mantle reservoir; DBAS, depleted basalt; TON, tonalite; UMR, unfractionated mantle reservoir; MIX, DMR-EMR mixture. DMR-EMR mixing and DBAS production are assumed to occur at 3.5 b.y. and DBAS metasomatism and tonalite production at 2.7 b.y.

Table 1. Parameters Used in Sm-Nd Isotopic Calculations

	Age (m.y.)	Sm (ppm)	Nd (ppm)	Sm/Nd	f (Sm/Nd)	ϵ_{Nd} (2700 m.y.)	ϵ_{Nd} (3500 m.y.)
DMR	4400	0.32	0.86	0.372	0.143	+6.1	+3.3
EMR	4400	8.75	35	0.25	-0.232	-9.9	-5.2
MIX	3500	0.40	1.2	0.345	0.060	+2.7	+1.5
DBAS	3500	1.7	5.2	0.327	0.116	+1.7	+1.5
EBAS	2700	5.0	20	0.250	-0.232	+1.7	-
TON	2700	3.5	24	0.146	-0.551	+1.7	-

MIX = 0.99DMR + 0.01EMR; DMR, EMR, depleted and enriched mantle; DBAS, EBAS, depleted and enriched basalt; TON, tonalite.

It is not until 2.7 b.y., however, that large volumes of tonalite are produced and rapidly added to the continental crust. Such rapid tonalite production may reflect catastrophic sinking of basaltic crust. Basalt that collects over sinks should resist recycling into the mantle because of its buoyancy and tendency to melt at relatively shallow depths. Perhaps 2.7 b.y. was a critical time in which geotherms in sinks decreased enough for the thickened basaltic crust to pass into the eclogite stability field, thus initiating catastrophic sinking into the mantle. In order to produce positive ϵ values yet high Nd contents in basalts that serve as tonalite sources, the DMR-EMR source mixture must include approximately half of the mass of the upper mantle.

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WAS CORE FORMATION VIOLENT ENOUGH TO HOMOGENIZE THE EARLY MANTLE?

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The dynamics of iron, its thermal state and its phase in the accreting earth probably played a major role in the earth's early thermal evolution. Plausible impact thermal histories (1,2) make it probable that pure iron was molten in the accreting earth after it was about 10% grown. Hence, iron eutectic alloys (FeS, FeO) certainly were (3). Additionally, the initial temperature of the core is an important constraint on the secular cooling of the early earth (4) and on the strength of the early geodynamo (5). Whether iron is solid or molten would influence geochemical equilibria in the upper and lower mantle (6); the mode of core formation, by spherical or near-spherical blobs (7,8), stalk-like instabilities (9), or something more catastrophic (10) would influence the partitioning of siderophiles between silicate and iron phases. Early descent of iron (during accretion) favors partitioning according to low-pressure phase equilibria, whereas late descent favors higher pressure. The later core formation occurs, the greater the heat pulse, due to the strong dependence of gravitational potential energy on planetary radius. The heat may homogenize the mantle if core formation is global; otherwise, heterogeneity of iron differentiation may leave some of the pre-archean mantle unaffected. The larger the chunks of proto-core (and hence smaller surface/volume ratios) the greater the heterogeneity.

The accreting earth probably had a thin surface layer of refractory (high-temperature) silicates (11), where heat transfer was dominated by impact-stirring (12,13), and possibly a central differentiated lithosphere (14). Farther out, the geotherm would follow an approximate r^2 -dependence (13) to a vigorously convecting region following the melting curve of the primary mantle silicate from the edge of the central lithosphere to the impact-stirred region. (15) suggests that a terrestrial magma ocean should be evidenced by an anorthositic crust and by the absence of ^3He . No such fractionated crust has been found, but primordial ^3He has been, both indicators suggesting a more complex history. However, heating by impacts is highly heterogeneous and core formation may not have been catastrophic enough to de-volatilize the mantle. Certainly, as earth began to retain heat, it could also retain volatiles which depressed melting temperatures (aiding core formation) and further enhanced mantle reservoirs after core formation. Continued bombardment over the next few 100 MY could have broken up an original refractory crust, leading to further fractionation to form the highly evolved crust characteristic of the oldest rocks found. Because of the temperature-dependence of viscosity, iron descent must be rapid on the accretion time scale (16).

We estimate the initial thermal and phase states of earth's core by examining the thermal evolution of spherical iron diapirs of various sizes descending in an accreting earth. Following (17,18), the thermal evolution of an iron body is the solution of a first-order ordinary differential equation. Heat transfer to or from the iron is conveniently expressed by the Nusselt number (Nu), which depends upon the descent velocity (v) and body radius (a) as parameterized by the Peclet number, $Pe = va/\kappa$, where κ is the thermal diffusivity of the mantle. (19) reviews transitions between several Nu/Pe relations as a function Pe range.

The core formation problem differs from silicate differentiation (17,18) because of the much greater differences in properties between the two chemical species. In the absence of heat sources, sinking iron tries to equilibrate with the surrounding silicates by boundary layer heat transfer to or from the silicate materials around it. In turn, surrounding silicates influence the iron and are themselves influenced by the strongly temperature-dependent viscosity of the mantle, providing a "self-regulation" (20). This feedback enables estimation of a minimum mantle viscosity for quasi-steady-state spherically-

symmetric accretion histories (16), and, thus, the limiting descent speeds of iron bodies.

Further, the thermal evolution of an iron blob will depend upon its descent speed. If Stokesian flow is assumed (Reynolds numbers are $\ll 1$ except for the largest bodies), then the descent speed depends upon body size, shape (see (20) for discussion), viscosity contrast with mantle (molten iron has free-slip surface boundary conditions that aid the flow), and gravity. Spherical iron diapirs test the limit of greatest heat retention. Various-sized bodies are chosen because constraints on sizes of impacting differentiated planetesimals or early instabilities are scanty.

Heat sources for iron bodies are (1) adiabatic heating, (2) radiogenic heating, (3) viscous dissipation and (4) latent heat. Given only adiabatic heating, spherical iron bodies with $a \ll 1$ km. equilibrate quickly to their surroundings. Larger bodies heat up, in spite of boundary layer heat loss. In the limit of no heat loss, bodies would have certain "potential temperatures" at certain depths. Large bodies below the geotherm eventually rise above it near the central lithosphere.

One possible radiogenic source for iron is K^{40} . (21) studied the electronic transition of K^{40} at high pressures. If K^{40} is siderophilic or chalcophilic, it may be responsible for long-term radiogenic heating as iron differentiates from the silicate matrix. (22) doubts such partitioning in experiments, and pressures necessary to cause the electronic transition in K^{40} (500 kb) are not reached statically within the mantle of an accreting planet, although dynamic pressures of 10 Mb may occur in large impacts (see Tables 2a,b in (30)). Following (23, p.140), the average K^{40} heat production 4.5 GY ago was only $\sim 1E-7$ ergs/g/sec. This would raise the temperature of iron containing all the K^{40} by only $2.5^\circ K/My$, certainly negligible over typical iron descent times (16).

Clapeyron curves for iron phase changes given by (24,25) allow iron's phase state to be followed from initial conditions. Sinking iron changes to more stable, denser crystal structures which may act as additional heat sources by releasing energy. A transition to liquid would absorb energy, helping to stabilize the temperature. Latent heats of transition depend upon Clapeyron slopes (dP/dT) and specific volume changes, Δv ($cm^3/mole$), given by (26) for several phases of iron.

The phase of iron determines descent velocities and the particular Nu/Pe number relation governing heat transfer, a highly non-linear process. While $v \propto a^2$, the increase in heat transfer from iron is slower because $Nu = 0.5v^{1/2}$ (liquid iron) and $Nu = v^{1/3}$ (solid iron). Ultimately, heat transfer is not as efficient as heat generation.

The assumption of spherical bodies leads to greatest heat retention, while long thin diapirs have the greatest viscous dissipation because their surface area is greatest. For flow in strongly temperature-dependent media, a body of arbitrary size descends more quickly than a body in isoviscous media because heat transfer from the body decreases the viscosity of wall rock in a surrounding boundary layer. Comparing a smaller and larger sphere, in a certain time Δt , heat will diffuse away from both spheres to the same distance, decreasing the surrounding viscosity. However, that distance is a greater fraction of the small sphere's size. Since a sphere advances by displacing its own volume, more of a small sphere's volume can pass through the lower-viscosity region than the larger's. (27) shows that heat transfer from a small sphere is significant; the sphere advances much faster than Stokes' law, but cools faster, too. The large sphere descends isoviscously in spite of the fact that there is a thin layer of lower viscosity around it. As (27) states, the thin deformation layer surrounding the large sphere "chokes." This, in a sense, is the limit of Tozer's (20) idea of "self-regulation." If a sudden influx of heat decreases the temperature-dependent viscosity, the greatest concentrations of heat produce the lowest viscosity regions, which may be narrow streams compared to

the global setting. That is, self-regulation may be valid only in a local setting because the nature of the temperature-dependent viscosity is to separate the effect into two different length scales: a small low-viscosity scale from a large isoviscous one.

The transformation of gravitational potential energy into heat on iron descent occurs through viscous dissipation. (23) examines viscous dissipation in Couette flow, similar to the boundary layer surrounding a large sphere. The dimensionless parameter governing viscous dissipation is $(1/2)PrE$, where Pr is the Prandtl number and E is the Eckert number. For $(1/2)PrE \ll 1$, viscous dissipation is negligible. However, Stokesian descent yields: $(1/2)PrE = (a^2 g^2 \Delta \rho^2) / (18 k \Delta T \nu)$. For $a=1$ km, $g=10^3$ cm/sec², $k=2.5 \times 10^8$, $\Delta \rho = \rho = 4$ g/cm³, $\nu=10^{18}$, and $\Delta T=10^3$ K, $(1/2)PrE=10^2$, viscous dissipation 2500 times higher than in the asthenosphere. Viscous dissipation arising from gravitational potential release along a column of mantle can be estimated by considering the energy balance between potential energy released and the increase in temperature of silicates and iron in contact with each other during the descent. Viscous dissipation is an important heat source for stalk-like diapirs or large spherical ones. (28) shows that bodies larger than a critical radius of 20 km can descend by gravitational energy-release melting the surrounding region. With other heat sources, we have shown critical sizes to be smaller by about a factor of 2-5. If such a body is in the shape of a long iron diapir, then viscous dissipation over its larger surface area will help to maintain its temperature so that the process becomes runaway. (13) calculates the plausible redistribution of gravitational potential energy of a planet differentiating into an iron core/silicate mantle when the solidus is reached. Because iron travels, on the average, farther than silicates, and hence, plausibly undergoes greater viscous dissipation, the greatest temperatures should occur near the core-mantle boundary. In addition, rising silicates adiabatically cool. Once again, great depths become hotter.

Finally, large deviatoric stresses from large iron bodies would rupture the cold primordial core (10). Deviatoric stresses of 10 kb or more might be common for iron stalks. These deviatoric stresses arise as much from heating as from density differences. From (23), thermal stresses are about $\Delta p = \alpha K \Delta T$, where K is the incompressibility. For $\alpha=3 \times 10^{-5}/^\circ\text{K}$, $K=1$ Mb ($=10^{12}$ dynes/cm²), and $\Delta T=300^\circ\text{K}$, $\Delta p=9$ kb. At high pressures, α decreases, but α of molten iron should be higher than that of solid iron (29).

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THE PROBLEM OF THE EARTH'S CO_2 CONTENT AND THE IRON CORE: G.W. DeVore, Dept. of Geology, Florida State University, Tallahassee, Fla. 32306

Paul Gast in 1972 (1) treated the problem of reduced iron in the Earth's core as the result of either its reduction by carbon from originally oxidized iron or the direct accretion of condensed metallic iron from the nebular dust cloud. The following is offered as an expansion of Gast's discussion.

The near absence of metallic iron and the presence of magnetite and FeS in the C-1 chondrites imply that metallic iron was a minor phase present during the accretion process that formed the C-1 chondrites. If the C-1 chondrites provided the bulk of the initial planetary growth materials, the carbon reduction model is favored.

The mass and radius of the Earth's core provide an average density from which can be estimated that 3.402×10^{25} moles of Fe^0 are in the core that could represent the reduction of 1.134×10^{25} moles of magnetite. If the core formed by carbon reduction of the magnetite, 2.268×10^{25} moles of CO_2 would be produced.

The total CO_2 in the Earth's crust is estimated by Ronov and Yaroshevsky (2) to be 9.07×10^{21} moles. If the Earth's mantle contains an average 100 ppm CO_2 , an additional 9.20×10^{21} moles may be present or 9.20×10^{22} moles of CO_2 if 1000 ppm of CO_2 are in the mantle.

The above estimates suggest that some 1240 to 227 times as much CO_2 may have been produced during the formation of the core than can be accounted for in the crust and mantle. This discrepancy taken to the extreme suggests either that:

1. The Earth has lost more than 99 percent of its initial CO_2 during early differentiation (this is highly unlikely) or:
2. The Earth has acquired some 90 percent of its present mass by the accretion of debris from previously reduced and differentiated but subsequently disrupted planetary bodies whereby the associated CO_2 would not be captured, or:
3. The C-1 chondrites represent only a trivial fraction of the initial accretion materials present in the nebular cloud or:

4. Condensed iron and anhydrous silicate phases were preferentially accreted during the initial formation of the planetary bodies as is favored in, for example, treatments by Clark et al. (3) and Anderson et al. (4).

An initial Earth some 10 percent of its present mass might retain CO₂ (Mars appears to retain CO₂). If so, the crustal inventory of CO₂ implies that only some 10 percent of the Earth's mass was accreted prior to its differentiation and iron reduction and that the rest of the mass of the Earth was acquired from second-hand recycled debris from disrupted differentiated planetary bodies.

The direct accretion of condensed Fe⁰ and anhydrous silicate eliminates the CO₂ problem. However, in a system where oxygen pressures are adequate for the formation of magnetite, it is difficult to explain how the bulk of the iron escaped oxidation unless the condensed iron had accreted into nodular masses before the magnetite formed. However, if nodules of Fe⁰ formed, it is difficult to imagine that these nodules were all removed from the nebular cloud so that the C-1 chondrites would not accrete these nodules. Accordingly, the model of the carbon reduction of magnetite retains some validity.

The crustal CO₂ content of the Earth transferred to the atmosphere would increase the mass of the atmosphere by 100 times and would be similar to that of Venus. If the accretion process of Venus followed that of the Earth, 90 percent of Venus' mass may have been derived also by accreting the differentiated debris of disrupted planets.

The discrepancy between the CO₂ produced during core formation and the CO₂ found in the crust and mantle would indicate a staggering amount of planetary growth by the meteorite impact model and a minor amount of initial growth directly from the nebular cloud.

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CONSIDERATIONS OF LARGE SCALE IMPACT AND THE EARLY EARTH. R. A. F. Grieve and

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Bodies which have preserved portions of their earliest crust indicate that large-scale impact cratering was an important process in early surface and upper crustal evolution. For example, large impact basins form the basic topographic, tectonic, and stratigraphic framework of the moon and impact was responsible for the characteristics of the second order gravity field and upper crustal seismic properties (1). The earth's crustal evolution during the first 800 my of its history is conjectural. The lack of a very early crust may indicate that thermal and mechanical instabilities resulting from intense mantle convection and/or bombardment inhibited crustal preservation. Whatever the case, the potential effects of large scale impact have to be considered in models of early earth evolution.

The number of impacts can be estimated from the cratering record in the lunar highlands, corrected for terrestrial impact conditions. With model approach velocities of $6-10 \text{ km s}^{-1}$ for the residual planetesimal swarm (2), it is estimated that the earth will collect 2-4 times as many bodies of the same mass as the moon and the resultant craters will be ~ 1.5 times larger, depending on the energy-diameter scaling relationship used. Assuming the size-frequency distribution of primary craters approximates $N \propto D^{-2}$, this translates to a correction factor of 4.5-9 for the number of craters of equivalent diameter D . From estimates of the cratering rate for the lunar highland (3) and the number of large lunar basins (4), the lower factor of 4.5 predicts ~ 200 basins with $D > 1000 \text{ km}$ on the earth in the period 4.6-3.9 by. This is considerably more than previous minimum estimates of $\sim 30-60$ basins with $D > 1000 \text{ km}$, which were based on relatively fast approach velocities of $15-20 \text{ km s}^{-1}$ (4, 5) and may be unrealistic as it requires a significant number of bodies in the residual planetesimal swarm, compared to the one or more of mass 10^{21} g required for basin formation on the moon (2).

Even for minimal estimates of the number of impacts, the values for additional exogenic energy and impact melt production are of the same magnitude as present-day processes of internal heat losses and island-arc volcanism, respectively (5). Among the effects of large-scale impact on an early terrestrial crust will be the formation of topography, enhancement of the sub-impact thermal gradients in the lithosphere and asthenosphere due to uplift of originally deep-seated materials, and the potential for eruption of basalt due to adiabatic decompression. Given that the earth has a relatively thin lithosphere, which is likely to have been even thinner during its early history, it is possible that large basin-sized impacts could bring the asthenosphere to the surface, leading to volcanism over large areas (4, 5).

Following impact, however, a number of processes act to modify large impact basins. These are: thermal contraction and subsidence from the loss of post-shock and uplift heat, topographic degradation by erosion and relaxation, and loading of the basin. Some of these effects, as they apply to basins on the moon, have been examined with the suggestion that lunar basins have appreciable effects, on timescales of 10^6-10^8 y , on the thermal, volcanic, and tectonic history in and around the basins (6, 7).

Preliminary models of the evolution of a large terrestrial impact basin have been undertaken. A 1000 km basin, with the extent of the sub-impact thermal anomaly constrained by uplift data from smaller terrestrial structures (8) and post-shock waste heat calculations (9), was formed in a model lithosphere 80 km thick with a thermal gradient of $20^\circ \text{C km}^{-1}$, overlying an asthenosphere with a thermal gradient of $0.5^\circ \text{C km}^{-1}$. Sub-impact temperatures were calculated as a function of time from finite difference solutions of the transient heat conduction equation in a cylindrical geometry (Fig. 1). Substantial conductive cooling occurs on a timescale of $\sim 10^7$ years. Consideration of the initial thermal anomaly and the melting behavior of anhydrous peridotite, suggests that $\sim 3 \times 10^6 \text{ km}^3$ of partially to totally molten mantle material will be brought to upper crustal and near surface depths. For more conservative gradients of $15^\circ \text{C km}^{-1}$ and $0.1^\circ \text{C km}^{-1}$, the initial thermal anomaly results in the immediate production of $\sim 8.5 \times 10^5 \text{ km}^3$ of at least partially melted peridotite. This is equivalent to an $\sim 1 \text{ km}$ thick layer of basalt within the basin. This is in addition to the production of impact melt, which for an approach velocity of 10 km s^{-1} is estimated to be of the same order of magnitude.

The initial topography of an 1000 km early earth basin, and how it changes with time, is not known. Some general statements, however, can be made. Isostatic adjustment will occur on a timescale of $2\mu/\rho g D$, where ρ and μ are the density and viscosity of mantle material underlying the basin and g is the acceleration of gravity. For $\mu = 10^{22}$ poise, isostatic rebound times are on the order of 10^6 years. On longer timescales the basin floor will subside due to conductive cooling and thermal contraction. Thermal subsidence will be concentrated near the center, where a maximum subsidence of $\sim 2 \text{ km}$ occurs after $\sim 10^8$ years (Fig. 2). In addition, the basin may be dynamically loaded by volcanics, sediments and water, which will cause further subsidence. For example, a sediment load with density $\rho_s = 2.7 \text{ g cm}^{-3}$ and $\rho = 3.3 \text{ g cm}^{-3}$ will cause a 1000 km basin with a surface topography of $\sim 3 \text{ km}$, a maximum based on depth-diameter ratios of terrestrial complex structures (8), to subside by an additional factor of 1.8; i.e., the original floor will subside to a depth of $\sim 5.4 \text{ km}$. This is in addition to thermal subsidence, which will contribute significantly to the thickness of sediments, particularly near the center.

These models do not account for several effects which will be important in determining the rate of heat transfer to the surface. First, loading of sediments into the basin will reduce the rate of heat loss and the rate of thermal subsidence. Subsidence due to loading will occur

on isostatic timescales ($\leq 10^4$ y) that are shorter than the time for conductive cooling of hot mantle beneath the basin (Fig. 1). Thus, basin-filling impact melt, breccia, post-impact volcanics, and sediments may be thermally metamorphosed and even partially melted to generate more silicic magmas. Second, solid-state convective upwelling beneath the basin will result in higher rates of heat transfer to the surface and to the basin fill. Convective flow will also cause radial spreading of the thermal anomaly resulting in the thermal subsidence being less concentrated near the basin center. Hotter mantle beneath the basin is less dense than surrounding mantle by an amount $\Delta\rho = -\rho\alpha\Delta T$, where α is the coefficient of thermal expansion and T is temperature which extends to a depth d . The resulting hydrostatic pressure difference $\Delta\rho gd$ must be balanced by a viscous stress $\mu w/d$ where w is the velocity of convective upwelling. Then, the time for upward convection of the thermal anomaly is $d/w = \mu/\rho\alpha g d$, while the conductive cooling time is d^2/κ , κ being the thermal diffusivity. For the initial thermal anomaly (Fig. 1a), $d \approx 50$ km and an average $\Delta T \approx 500^\circ\text{C}$. With $\mu = 10^{22}$ poise and appropriate values of the other parameters, the time for convective upwelling is shorter than that for conductive cooling by a factor of 10^2 . This is a minimum estimate because the mantle viscosity of an early, hotter earth is likely to be less than a value typical of the present earth. Thus convective upwelling beneath the basin should significantly enhance heat transfer to the basin fill and influence both the rate and radial distribution of thermal subsidence.

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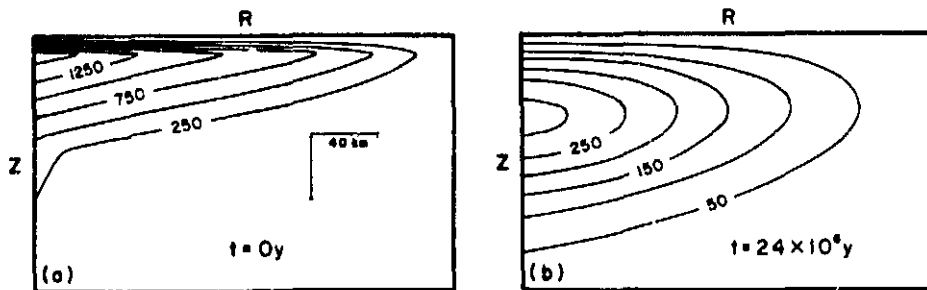


Figure 1. Vertical section of thermal anomaly beneath model 1000 km impact basin. R is radial distance and Z depth. Contours are for temperatures in excess of pre-impact thermal gradient. (a) Initial post-impact anomaly; (b) After 24×10^6 y.

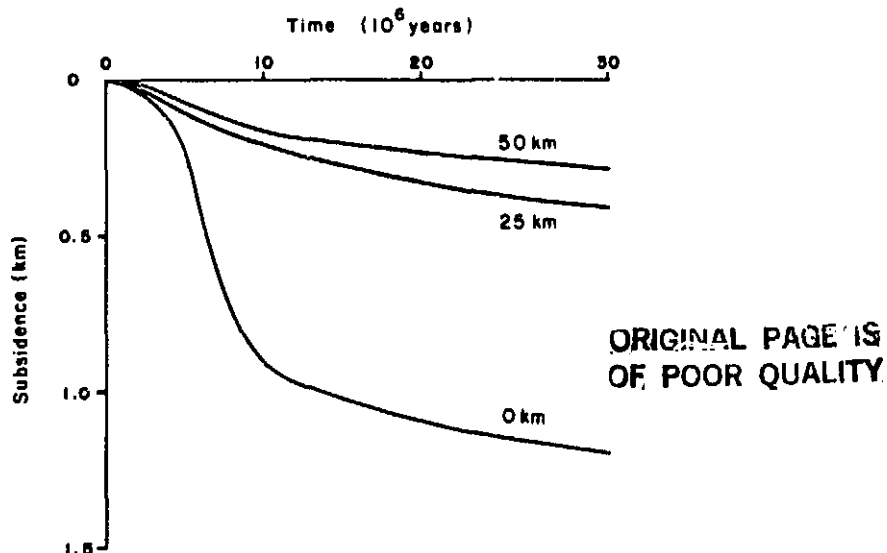


Figure 2. Thermal subsidence at various distances from the center of 1000 km basin, based on post-impact conductive cooling model, as a function of time.

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Hydrothermal alteration in the 3.5 b.y. old
Onverwacht Group of South Africa

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K-Ar studies of authigenic and alteration phases from the 3.5 b.y. old Onverwacht Group of South Africa delineate hydrothermal metamorphism that ended 3.3 b.y. years ago. As shown in Fig. 1 a whole rock K-Ar errochron from analysis of barite, dolomite, chert, and serpentinite (komatiite) gives an age of 3.3 b.y. with an intercept of $1,678 \pm 103$. The 3.3 b.y. age for the metamorphism in the Onverwacht has been confirmed by the ^{40}Ar - ^{36}Ar step-wise heating experiments of komatiites and basaltic komatiites from the Onverwacht Group (1). In addition, the errochron suggests all the phases studied equilibrated with a reservoir of hydrothermal argon with relatively uniform isotopic composition.

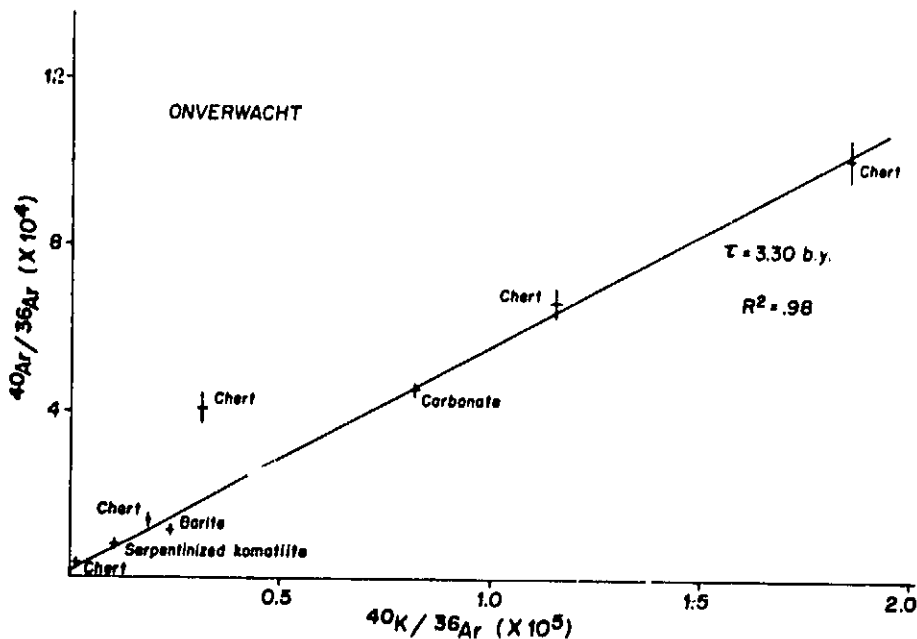


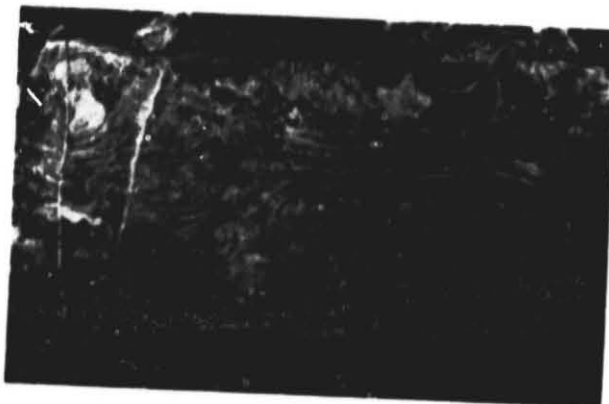
Figure 1: A 3.3 b.y. K-Ar isochron of hydrothermal phases from the Onverwacht Group, Barvarton Mountain Land, South Africa (from Reference 2).

R. Hart & M. de Wit

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The concept of hydrothermal activity in the Onverwacht is supported by the discovery by M.J. de Wit of fossilized hydrothermal vent structures as shown in Fig. 2 (2).

Figure 2: Silicified mud pool structures thought to have formed by hydrothermal emanations (from Reference 2).



In addition, the degree of hydration of Onverwacht rocks is indistinguishable from that of rocks dredged from present day ocean ridge spreading centers. This is shown in Figure 3, a plot of H_2O vs. MgO .

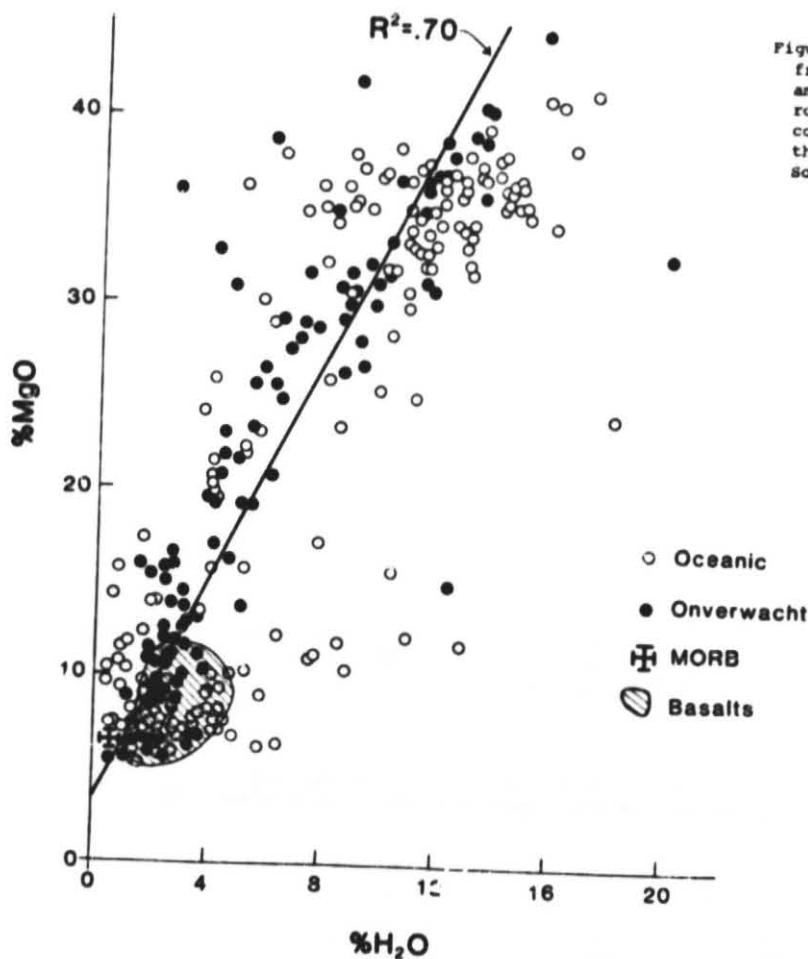


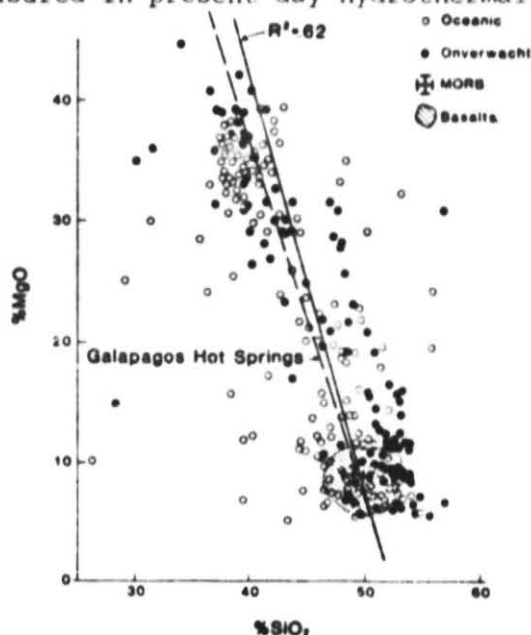
Figure 3: A plot of % MgO vs H_2O for fresh basalts, greenschist facies, amphibolite facies, and ultramafic rocks recovered from the ocean floor compared to the same rock types from the archaean Onverwacht Group of South Africa.

R. Hart & M. de Wit

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The hydrothermal alteration in the Onverwacht is characterized by extensive mobilization of and metasomatism by silica resulting in extensive silicification of cross bedded shales, accretionary tuffs, carbonates, barites, iron rich sediments and stromatolites. As shown in Fig. 4, leaching of silica from the volcanic rocks by hydrothermal solutions was accompanied by addition of magnesium. Indeed, the ratio of silica removal to magnesium addition is indistinguishable from that measured in present day hydrothermal springs at the Galapagos Rift (3).

Figure 4: Plot of % Mg vs SiO_2 for basalts, greenschist facies, amphibolite facies and ultramafic rocks recovered from the ocean floor compared to those from the archaean Onverwacht group. The best fit regression line for all data gives $\text{SiO}_2/\text{MgO} = .28$. The ratio of the SiO_2 flux exiting to the flux of MgO entering at the Galapagos hot springs vents is .24 (3). This suggests the MgO - SiO_2 trend shown here may have been produced by hydrothermal leaching of Si and addition of Mg.



Carbonate stromatolites as shown in Fig. 5 are stratigraphically associated with the fossil hydrothermal vents and display possible microfossils suggesting biological activity was associated with these vent systems 3.3 b.y. ago (2).

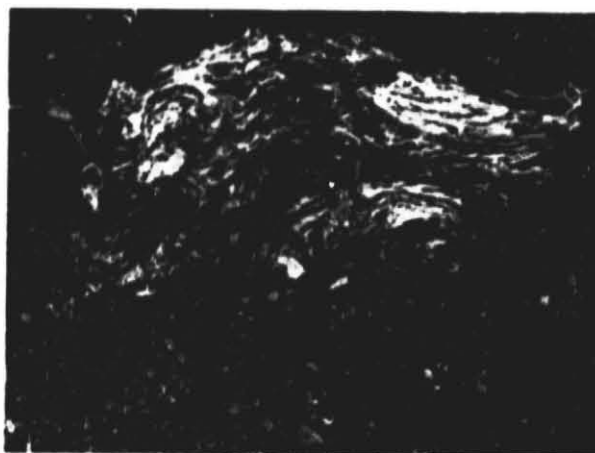


Figure 5: Stromatolites in carbonate from the Onverwacht Group, Barberton Mountain Land, South Africa (from Reference 2).

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The degassing history of the earth;
noble gas studies of Archaean cherts
and zero age glassy submarine basalts.

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Recent noble gas studies suggests the earth's atmosphere outgassed from the earth's upper mantle synchronous with sea floor spreading, ocean ridge hydrothermal activity and the formation of continents by partial melting in subduction zones. This process is illustrated in Figure 1 (from reference 1). The evidence for formation of the atmosphere by outgassing of the mantle is the presence of radionuclides ^4He , ^{40}Ar and ^{136}Xe in the atmosphere that were produced from ^{40}K , U and Th in the mantle. Allègre et al. (2) suggest the mean age of the atmosphere is 4.4 b.y. However, 1 b.y. - 3 b.y. is required to produce the ^{40}Ar in the atmosphere by decay of ^{40}K in the mantle. As shown in Figure 2, concordant mean ages for the cessation of degassing are in the range 1.2 - 1.8 b.y. ago as derived from the amount of ^{40}Ar in the atmosphere as well as the concentration of ^{40}Ar in glassy MORB (3). The He and Ar isotopic compositions of glassy submarine basalts as shown in Figure 3, require a degassed mantle (100 times depleted in ^{36}Ar) and non-degassed mantle. The relatively uniform $^3\text{He}/^4\text{He}$ ratio in MORB (2) suggests the outgassing of ^3He was uniform in aerial and temporal extent. Less than 5% of present day ^4He in the mantle is being degassed. These observations allow fitting of closed system growth curves to $^{40}\text{Ar}/^{36}\text{Ar}$ and $^3\text{He}/^4\text{He}$ ratios of the mantle reservoirs as shown in Figure 4. This two stage model with cessation of degassing at 1.8 b.y. ago yields similar results for fractionation of the $^3\text{He}/^{36}\text{Ar}$ by degassing of the MORB source as required by the fit of mixing lines in Figure 3. Cessation of degassing in the interval 1.2 - 1.8 b.y. with a possible peak of hydrothermal activity 2.5 b.y. ago is also suggested by the initial $^{40}\text{Ar}/^{36}\text{Ar}$ of hydrothermal authigenic cherts as shown in Figure 5. Thus, the cessation of degassing is synchronous with the increase in $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of sea water produced by the curtailment of hydrothermal activity and the increase in stream supply of radiogenic strontium from continental masses (4).

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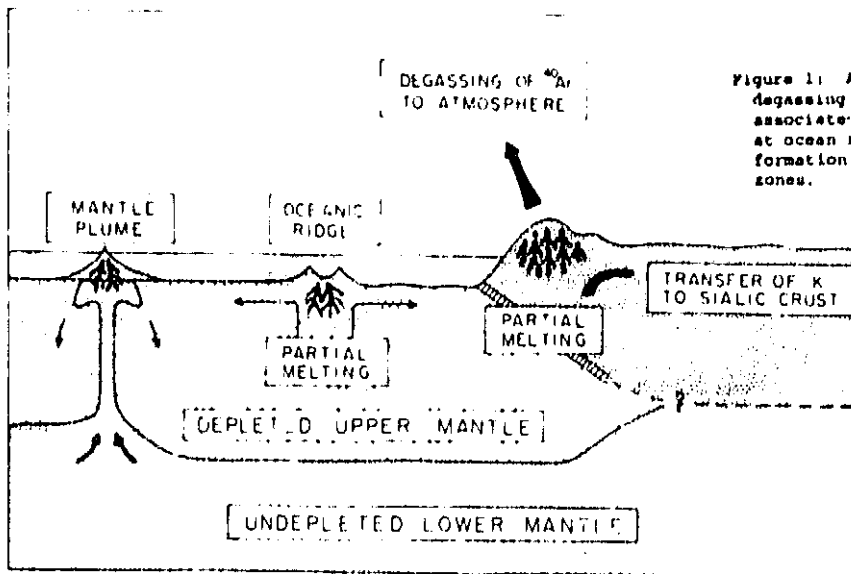
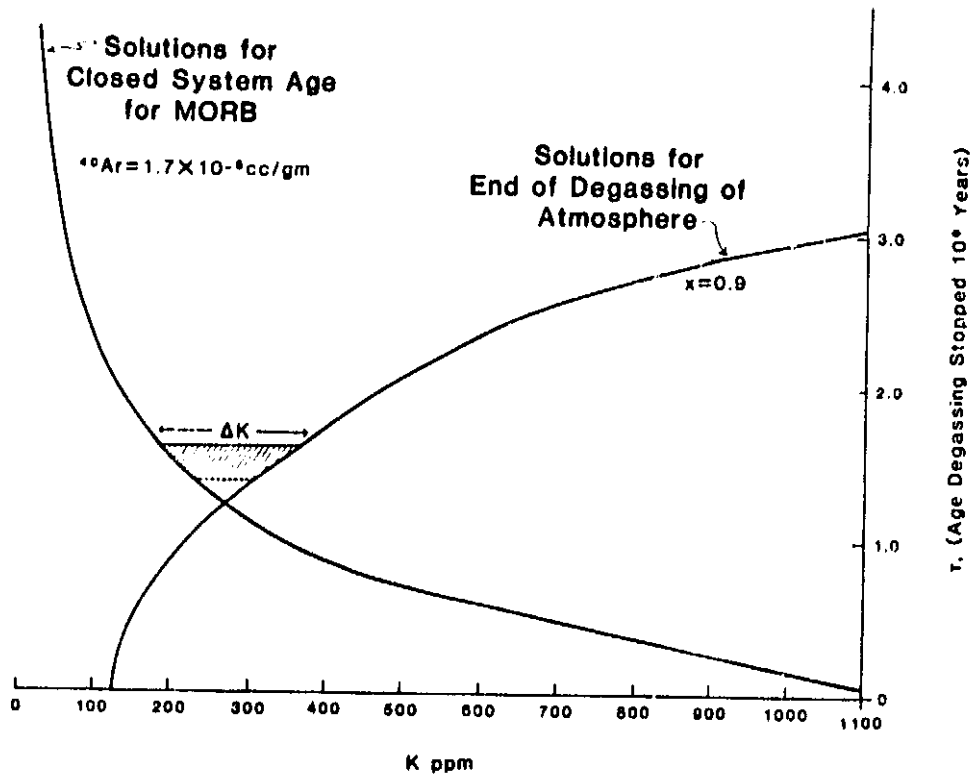


Figure 1: A schematic representation of degassing of ^{40}Ar to the atmosphere associated with hydrothermal activity at ocean ridge centers and continent formation by volcanism at subduction zones.

Figure 2: Convergent solutions for τ , the mean age of the cessation of degassing, as a function of potassium content of the earth is shown. Solutions for cessation of degassing of atmosphere assume the ^{40}Ar content of 21.79×10^{21} cc to have degassed from 90 percent of the mantle ($x = .9$). Solutions for beginning of the retention of ^{40}Ar in MORB assume an ^{40}Ar concentration for MORB of 1.7×10^{-6} cc/gm. ΔK symbolizes the difference in the K concentration between the degassed and non-degassed mantle. The top of the shaded area corresponds to 90% of the mantle degassed while the bottom represents 50%.



Hart, R. & Hogan, L.

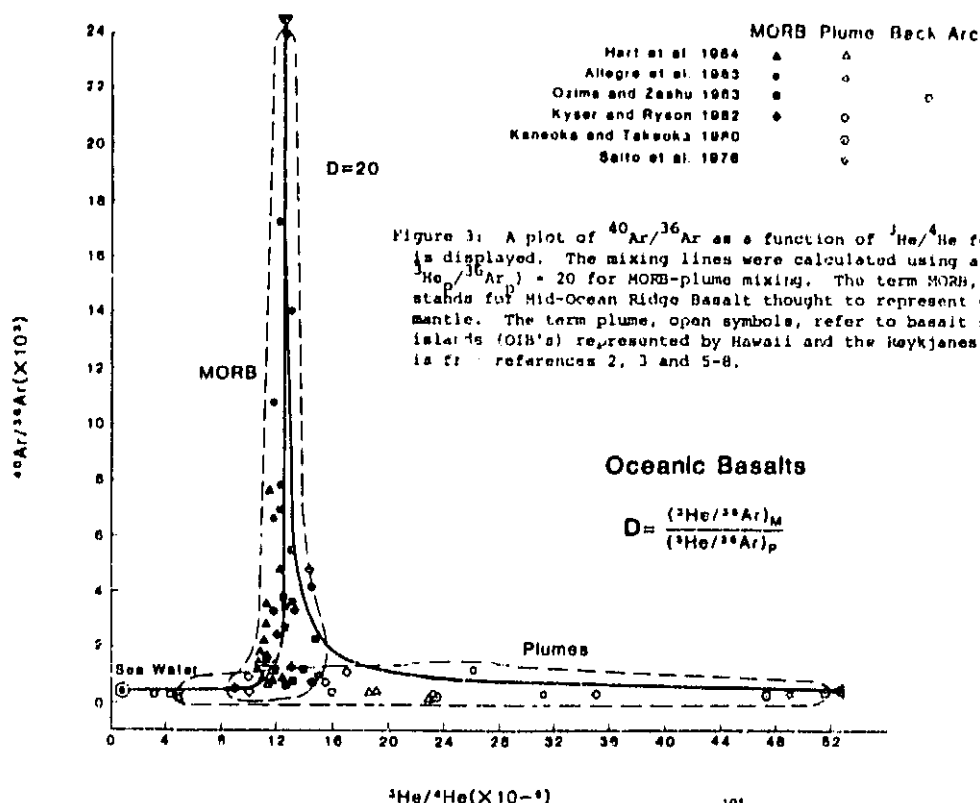


Figure 4: Two stage growth curves for the production of the $^{40}\text{Ar}/^{36}\text{Ar}$ and $^3\text{He}/^4\text{He}$ ratios in non-degassed mantle and in MORB are shown. It is assumed that the MORB source degassed up until 1.9 b.y. ago and then was virtually closed to argon and helium. This assumption is supported by the observation that 95% of ^4He production is currently retained in the mantle and the uniform $^3\text{He}/^4\text{He}$ ratio of MORB. The value for $D = 14.5$ is derived from the ratio $(\mu_A/\mu_H)_M / (\mu_A/\mu_H)_P$ assumes no change of the K/U ratio during the depletion.

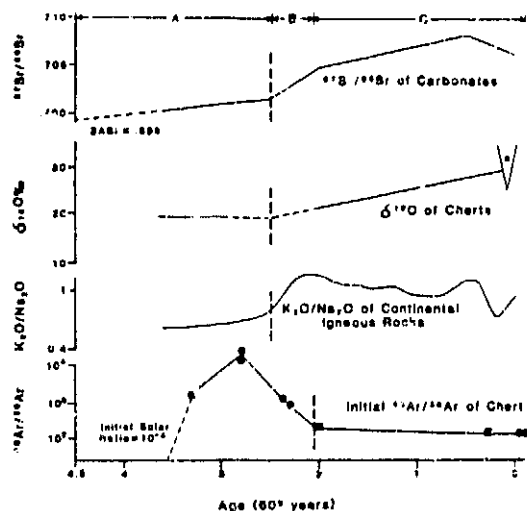
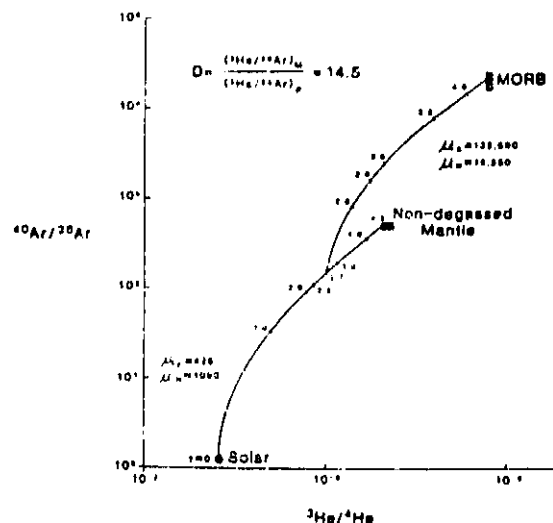


Figure 5: A plot of $^{87}\text{Sr}/^{86}\text{Sr}$ in carbonates (reference 4) $\delta^{18}\text{O}$ in cherts (reference 9) $\text{K}_2\text{O}/\text{Na}_2\text{O}$ in continental rocks and initial $^{40}\text{Ar}/^{36}\text{Ar}$ ratio in cherts with age. These plots suggest sea water was dominated by hydrothermal input for the first 2.0-2.5 b.y. of earth history, as continental runoff became dominant the $^{87}\text{Sr}/^{86}\text{Sr}$ and $\delta^{18}\text{O}$ increased. In a similar fashion mixing of atmospheric argon with hydrothermal argon caused a decrease in the $^{40}\text{Ar}/^{36}\text{Ar}$ ratio of sea water.

THE 3.5 B. Y. OLD ONVERWACHT GROUP:
A REMNANT OF ANCIENT OCEANIC CRUST

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The interaction between seawater and submarine volcanic rock has had important consequences for the chemistry of the ocean during the Phanerozoic. Processes of hydrothermal chemical exchange and mineralogical alteration have been documented extensively for Phanerozoic ophiolites and oceanic crust, but evidence for similar processes occurring early in Earth's history has been controversial. It is now generally accepted that most of the mineral wealth of the Canadian Shield is the result of hydrothermal activity during the later Archaean and the early Proterozoic. However, it has been difficult to document occurrences of hydrothermal alteration in the early Archaean. Most extant terranes have been regionally metamorphosed to the amphibolite and granulite facies, so that their precursor lithologies and structures are not readily determinable. However, the 3.5 b. y. old supracrustal rocks of the Barberton Mountain Land, South Africa, have not been subjected to high grade regional metamorphism, and therefore there was reason to hope that a laboratory investigation might reveal the extent to which these rocks had been exposed to subseafloor hydrothermal activity. Hart and de Wit (this volume) describe bulk geochemical evidence from the entire suite as well as field evidence which support the concept of hydrothermal activity in the Barberton Mountain Land. This paper briefly discusses mineralogical and textural features which unequivocally mark it as a submarine sequence emplaced in a mid-ocean ridge/fracture zone or back arc/fracture zone environment.

As a general rule, there are two features which distinguish an ocean floor metamorphic sequence from a regionally metamorphosed sequence: the first is that ocean floor metamorphism entails very steep thermal gradients so that distinct facies changes can occur over very short distances; the second is the preservation of primary igneous textures in untectonized amphibolite facies rocks. Therefore, in order to identify the metamorphic environment of mafic and ultramafic rocks now found on continents, one must consider not only the mineralogical assemblages but also the lithological associations and textural character, features which will indicate the scale of the metamorphic gradient.

The classical greenschist facies of regional metamorphism was defined from the study of rocks on continents. Application of this facies concept to altered mafic and ultramafic rocks is perilous because the mineralogy of the classical greenschist facies is indistinguishable from the greenschist mineralogy found in rocks dredged from the ocean floor (1,2,3). Both types of rocks will contain the same index minerals: chlorite, actinolite, epidote, albite. Thus, differentiating between metamorphic environments solely on the basis of the greenschist facies mineralogy of mafic and ultramafic rocks is not possible.

The classical amphibolite facies mineralogy of regional metamorphism is characterized by the presence of hornblende and andesine (1). Furthermore, the process of progressive regional metamorphism involves the loss of fluids and the complete recrystallization of the greenschist mineralogy. Primary textures are almost never preserved. Instead, the textures of the amphibolite facies of regional metamorphism reflect the operant tectonic forces: for ex-

ample, preferred orientation, lineation, foliation, gneissic layering, and annealing in crystals (1,2). A further characteristic of regional metamorphism is the field recognition of metamorphic isograds using mineral assemblages. Appearances and disappearances of mineral phases defined zones of large areal extent, and the scale of the dynamothermal gradient is measured in kilometers.

While amphibolite facies rocks with dominantly hornblende-plagioclase mineralogy have been dredged from the ocean floor, they are relatively rare compared to spilites and serpentinites and generally contain numerous small fractures lined with retrograde hydrous phases such as zeolite, epidote, and chlorite (3). Metagabbros described by Bonatti et al. (4) had a transitional greenschist-amphibolite facies mineralogy with both hornblende and tremolite-actinolite being found in the same rocks. In addition, these metagabbros were associated with both greenschist facies metabasalts and unmetamorphosed glassy basalts. Such closely-associated lithologies would not occur in a regionally metamorphosed terrane. Another significant feature of metagabbros from the ocean floor is their retention of primary igneous textures. While rocks from fault scarps along axial rift zones have been crushed and sheared, many ocean floor amphibolites retain the subhedral-granular, poikilitic, or subophitic textures of their primary mineralogy. The same textural preservation is observed in unshaped lavas and cumulates.

The Onverwacht Group is the lowest division of the Swaziland Supergroup of the Barberton Mountain Land. Tankard et al. (5) describe it as "primordial ocean-type crust". It is a sequence of altered mafic and ultramafic lavas, tuffs, and cherts, intrusive gabbros, and bodies of layered ultramafic and mafic cumulates (see ref. 5,7). Each formation in the Onverwacht Group contains rocks of at least two metamorphic facies (zeolite/greenschist; greenschist/amphibolite) and some contain all three (zeolite/greenschist/amphibolite). The Swaziland Supergroup has many features of an ophiolite/flysch/molasse terrane, but identification of it as an Archaean ophiolite is controversial because of the presence of ultramafic lavas (komatiites) and the lack of a readily-identifiable sheeted dike complex.

In areas of structural disturbance, such as the boundaries near plutonic contacts or along internal fault zones, Barberton rocks show the expected textural disturbances. However, rocks from unfaulted sections of the interior retain primary textures to an astonishing degree. Pillow cross-sections retain textural evidence of submarine crystallization. Some basalts show concentric zoning in color and texture which in Phanerozoic ocean floor rocks are characteristic of low temperature alteration. Variolitic textures, which develop from undercooling in basic lavas (2), are also preserved, though the plagioclase is now albite. Alteration mineralogies are frequently pseudomorphic after primary mineralogies: serpentine after olivine, amphibole after pyroxene, chlorite after devitrified glass, etc. Textures are so well preserved that it is the type area for spinifex textured komatiites (10). In thin section, the nature of the precursor is often determinable on the basis of texture alone.

Another characteristic of these rocks is the presence of fractures, ranging in size from tenths of a millimeter to several centimeters. These fractures are filled with a variety of minerals and mineral assemblages. Epidote and clinozoisite are very common in lavas and gabbros. Quartz and calcite fill both fractures and vugs in mafic lavas. Serpentinized dunites and pyroxenites have fractures filled with talc as well as serpentine and magnetite. SEM photos of crushed samples show that the serpentines from the ultramafic and mafic cumulates have the platy habit of antigorite.

Hoffman, S.

Alteration in the Onverwacht was not isochemical as originally proposed (10). Rather, the rocks record a complex history of elemental mobility. Many of the ultramafic rocks of the Stolzberg Ultramafic Complex, which had little or no primary calcium as determined from the relict mineralogy, contain calcic alteration minerals such as tremolite and leucoxene. Ultramafic bodies in the Barberton Mountain Land also contain rodingites, intensely metasomatized rocks now consisting of extremely calcic hydrous alteration minerals. A rodingite from the Stolzberg Ultramafic Complex with relict hypersthene now contains over thirty weight percent CaO. Microprobe analyses of this rock determined that it contained four calcic garnet compositions (grossular, hydrogrossular, grossular-andradite, and titanium andradite) as well as tremolite, zoisite, apatite, jurupaite, and diopside.

While the mineralogies of the rocks from the Barberton do not rule out a regional greenschist facies event, recent analyses by Lopez-Martinez et al. (8) indicate that alteration of the sequence occurred very shortly after eruption of the lavas. Also, there is no mineralogical evidence for large-scale regional compression. The persistence of antigorite, which readily converts to chrysotile during regional greenschist facies metamorphism, suggests that any regional metamorphic event affecting the Barberton could not have been very pervasive.

At the present time, the only environment where the mineralogical and lithological associations described here are known to form is in the deep ocean near active rifts (amphibolitic metabasalts and metagabbros) or in fracture zones (metabasalts, metagabbros, serpentinites, rodingites). In the Equatorial Atlantic, rodingites were only dredged from fracture zones and only found in association with serpentinites (3,9). The estimated temperatures of formation of Equatorial Atlantic rocks (9) overlap those estimated for the Onverwacht rocks on the basis of oxygen isotope studies. The presence or absence of a sheeted dike complex need not be regarded as diagnostic of oceanic crustal origin because not all oceanic sequences appear to have them (3). Overall, the correlations between metamorphic rocks from the Equatorial Atlantic (3,4,9) and those from the 3.5 b. y. old Onverwacht Group are striking. The data described here and elsewhere (5,6,7; Hart and de Wit, this volume) indicate that the Onverwacht Group is not "primordial ocean-type crust" (5) but is Archaean oceanic crust. Further study of this sequence is likely to greatly enhance our understanding of Archaean oceanic and crustal evolution.

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SUBMARINE HYDROTHERMAL ENVIRONMENTS
AS SITES FOR THE ORIGIN AND EVOLUTION OF LIFE

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There are currently several hypotheses as to how and where life originated and evolved. Assuming that life did arise on Earth, the important question then is what environment provided the chemical and physical conditions for the reactions leading to organic compounds, self-replicating and information-carrying macromolecules, and eventually organisms and communities. In order to have met all of the conditions necessary for these reactions to have occurred, there had to have been on the early Earth either a multiplicity of closely-linked environments or a single environment where gradients existed in the physical and chemical conditions. Such an environment would have provided an almost infinite set of conditions and variables, as well as combinations of conditions. There is only one environment in the Phanerozoic which fulfills the requirements of pre-biologic chemistry and physics: submarine hydrothermal vents. Table 1 lists the properties of vents as well as the environmental conditions within and around them. These properties and conditions are also known prerequisites for the origin and development of organisms. In a previous paper, we proposed that submarine hydrothermal systems possessed all of the necessary criteria to have been sites for the origin and evolution of life (1). We based our hypothesis on three lines of evidence: the properties of present-day hydrothermal environments; the lithological associations of the most ancient fossils; and the laboratory synthesis of amino acids and complex organic compounds using thermal energy sources.

Recent models for the formation of the atmosphere and hydrosphere (2,3) indicate that an ocean had formed as long ago as 4.2 b.y. At the time of the ocean's formation, the Earth had an oxidizing atmosphere consisting mainly of CO_2 (2,3). Before the formation of continents, the ocean covered the earth and all volcanism was submarine, resulting in extensive hydrothermal activity. As a consequence of rapid hydrothermal circulation, this early ocean was anoxic (4) and was probably hotter than 50° to 70°C (5) due to the high partial pressure of CO_2 in the atmosphere. It is in association with these ancient hydrothermal environments that we postulate life to have originated and evolved into the complex fossil microbial communities observed in rocks from the early Archaean (6,7,ref. cited in 1,8).

Unicellular fossils and stromatolites have been found in the 3.5 b.y. old Warrawoona Group (Pilbara Block, Western Australia) (6) and in the 3.5 to 3.3 b. y. old Swaziland Sequence (Barberton Mountain Land, South Africa) (7). The fossil-bearing strata in both localities are part of submarine sequences which include lava flows, pillow basalts, and chemical sediments which were probably derived from hydrothermal sources (ref. cited in 1). Complex microorganisms having morphological similarities to these Archaean microfossil assemblages have been found in modern hydrothermal environments (8).

The precursors to complex organic compounds and macromolecules, such as CH_4 , CO , H_2 , H_2S , and reduced metals, are released in present-day hydrothermal systems as a result of water-rock interactions at high temperatures. Our model for abiotic synthesis in hydrothermal systems proposes that chemical species such as these react in a step-wise series to produce successively more complex organic molecules. Energy to drive the reactions is derived in part

Hoffman, S., and Baross, J.

from the heat of crystallization of erupted and emplaced igneous material and in part from the heat of hydration released when these rocks are hydrothermally serpentinized, chloritized, uraltized, or spilitized. The calculations of Fyfe (9) indicate that about 3×10^8 calories are released by the hydration of each square centimeter column of oceanic crust seven kilometers in thickness. For a lateral spreading rate of six centimeters per year, heat of hydration alone will provide roughly 18×10^8 calories per centimeter of ridge length per year.

In 1980, at the time of our original paper (10), the concept of Archaean hydrothermal systems was controversial, particularly with regard to the nature of the strata in which the oldest fossils have been found. More recent work (see Hart and de Wit, Hoffman, this volume; ref. 11) has demonstrated that not only is the concept of pre-Archaean and Archaean hydrothermal activity valid but also that it was probably much more extensive than during the present day. Abbott and Hoffman (11) showed that the major geological features of extant Archaean terranes can be explained by a model in which the overall rate of seafloor creation and subduction has steadily declined over the history of the Earth. Their model proposes that Archaean ridge length was 3 to 4 times greater 3.5 billion years ago than at the present time. Extrapolation back to 4.2 billion years ago suggests that then the total oceanic ridge length was as much as 5 times greater than at present.

The present ridge length is 56,000 kilometers (5.6×10^9 centimeters). A pre-Archaean ridge five times longer than the present would therefore have been roughly 280,000 kilometers (2.8×10^{10} centimeters). If we assume an average pre-Archaean spreading rate of 6 centimeters per year (probably a conservative estimate), and assuming an estimated 18×10^8 calories of thermal energy (heat of hydration) are released per centimeter of present-day ridge, a pre-Archaean mid-ocean ridge system would have potentially generated 5×10^{19} calories of thermal energy through the hydrothermal cooling of freshly emplaced oceanic crust per year. This strongly suggests that submarine hydrothermal activity on the early earth was the most dominant and most dependable source of energy to drive the reactions necessary for abiotic synthesis. Furthermore, on an Earth without continents, the ocean had to have been more shallow, assuming an equivalent volume of ocean water. Therefore, the mid-ocean ridge depth on the early Earth was probably less than 1000 meters. In addition, the evidence for the degassing history of the earth (see Hart and Hogan, this volume) in general suggests that the overall rate of mantle degassing was substantially higher prior to 3.5 b. y. ago, so that the supply of chemical precursors for abiotic synthesis was possibly greater.

We have proposed that life formed and evolved in hydrothermal environments. This hypothesis is plausible in terms of the tectonic, paleontological, and degassing history of the Earth. Submarine hydrothermal vents are the only contemporary geological environment which may truly be called primeval and which today continue to be a major source of gases and dissolved elements to the ocean. The microbial assemblages in present-day hydrothermal systems therefore could be living analogues of the earliest microbial communities to develop on Earth.

TABLE 1. Hydrothermal Vents and Associated Oceanic Gradients.

Properties and conditions important for the study of the origin
of life and related evolutionary processes.

ENVIRONMENT	PROPERTIES, CONDITIONS, VARIABLES
I. Submarine Hydrothermal Vents (open & geologically continuous)	<ol style="list-style-type: none"> 1. Temperature (~500°C to ~2°C in the present ocean; ~500°C to ?°C in the Archaean ocean) 2. Pressure (? - dependent upon depth of the ridge) 3. Concentrations of reduced metals (microbial energy sources), trace metals, & gases (H₂S, CH₄, H₂, CO, N₂, NH₃, etc.) 4. Organic compounds (?) 5. Condensing agents (CN, CH₂O, polyphosphates, etc.) 6. pH 7. Reaction surfaces (clays, zeolites)
II. Surface Waters/ Archaean Ocean	<ol style="list-style-type: none"> 1. Light 2. Temperature (?) 3. Pressure (? - dependent upon partial pressure of atmospheric CO₂) 4. Oxygen (? from photolysis) 5. N₂O, NO, other biologically important gases 6. Carbon and energy sources (see I-3, above)
III. Other Associated Environments	<ol style="list-style-type: none"> 1. Submarine rock/water interfaces 2. Subaerial rock/water interfaces 3. Atmosphere/ocean interfaces

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NOBLE GASES AND THE EARLY HISTORY OF THE EARTH: INAPPROPRIATE PARADIGMS AND ASSUMPTIONS INHIBIT RESEARCH AND COMMUNICATION. G.R. Huss and E.C. Alexander, Jr., Dept. Geology and Geophysics, Univ. of Minn., Minneapolis, MN 55455

Noble gases are attractive tracers of volatiles through the Earth's evolution due to their apparent chemical inertness, presumed conservative nature and links to the important heat producing elements. Numerous models of the Earth's volatile evolution have been proposed, debated and ultimately abandoned. Many of the obsolete models included terms which are recycled into later models. In many cases not only the terms, but portions of the associated models are carried forward. A number of familiar terms were originally defined in meteorite or lunar studies and carry extensive paradigms which may not be entirely appropriate to terrestrial problems. Widespread use of common terms, which mean very different things to different individual researchers, contributes much to the confusion in terrestrial noble gas studies. In addition, simplifying assumptions are often used even when they are demonstrably inconsistent with physical reality, obscuring the significance of experimental results and leading to confusing models.

The familiar straw man of a primary versus secondary atmosphere is one example of the type of confusion that can occur in models of volatile evolution. Primary atmospheres were originally defined in terms of a direct, gravitational capture of gases from a solar nebula. Secondary atmospheres were defined as those produced by the accumulation of volatiles from the degassing of a planet after the loss of a primary atmosphere. The additional connotation of a long time scale for the degassing was quickly included in the conventional usage. Most, but not all workers, have abandoned primary atmospheric models and, therefore, the necessity of the loss of a primary atmosphere in secondary atmospheric models. Catastrophic early degassing models propose the major release of volatiles occurred during accretion or initial differentiation of the planet. Within these models, the initially degassed volatiles become the primary atmosphere and subsequent physical and chemical processes produce the current, secondary atmosphere. All of the above usages occur in current models.

The concept of degassing itself presents problems. No significant data exists on the distribution coefficients of volatiles (or noble gases in particular) between solid and liquid phases for mantle temperatures and pressures. In the absence of experimentally determined distribution coefficients, modelers make a series of assumptions. For example, it is commonly assumed that volatiles and the noble gases are efficiently partitioned into a melt phase and that volatiles and noble gases behave similarly enough to use noble gases as tracers. The noble gases are assumed to be chemically inert. And they are assumed to diffuse at similar enough rates to allow them to be treated as a coherent group. These assumptions generally arise from laboratory experience on the distribution of noble gases between solid and gas phases at one atmosphere of lower pressures. The apparently contradictory pictures of the mantle presented by data on the various noble gases indicate that one or more of these assumptions is significantly incorrect. It is by no means obvious, for example, that helium can tell a liquid from a solid at mantle temperatures and pressures.

Equilibrium concepts are widely used when attempting to predict noble gas or volatile behavior in the solar nebula, in large part because they yield quantitative results. There is abundant evidence, however, that thermal, chemical, and mixing equilibria were not maintained in the solar nebula

Huss, G.R. et al.

and there is every reason to expect significant kinetic effects. The enormous thermal disequilibrium, astronomically observable between solid and gas phases of nebulae, may well prove to be the key factor in determining the noble gas abundance in the solids from which the planet forms.

Much of the discussion of the Earth's noble gas inventory is conducted in terms of solar and planetary components defined from meteorites. The Earth's noble gases fit neither component isotopically or elementally. Hundreds of pages of scientific literature have been expended attempting to relate solar, planetary and terrestrial noble gases to one another and/or to some yet more primordial composition. There is abundant evidence that the meteoritic planetary composition is not a unique component but is made up of several separable components. Clearly a single isotopically homogeneous primordial composition is not consistent with a wide variety of isotopic, chronologic, and chemical evidence. However, a single primordial composition is such a seductively useful concept, it winds up permeating any model that does not explicitly rule it out.

We must develop a new set of paradigms embodying our present knowledge to guide future investigations. The concepts in the new paradigms will hopefully be clearly divorced, in both the writers' and readers' mind, from their historical antecedents. There are several important areas for future research. The first is to measure the elemental and isotopic compositions of the five noble gases in a large number of terrestrial materials, thus better defining the composition and distribution of terrestrial noble gases. To a remarkable degree these measurements remain to be done. Second, determinations of relative diffusive behavior, chemical behavior, and the distribution between solid and melt of noble gases under mantle conditions are urgently needed. Finally, disequilibrium behavior in the nebula should be investigated, including the behavior of plasmas and the consideration of possible cryotrapping on cold nebular solids.

CONSTRAINTS ON THE DEVELOPMENT OF THE EARLY CONTINENTAL CRUST FROM ISOTOPIC DATA, S. B. Jacobsen, Dept. of Geological Sciences, Harvard University, Cambridge, MA 02138

In principle many aspects of the origin and development of the early (4.5-3.8 AE) continental crust can be addressed by radiogenic isotope and trace element studies. The most important ones are (i) at what time did the earliest continental crust form?, (ii) what was its composition?, (iii) by what processes did it grow and by what processes was it destroyed, (iv) what were the rates of production and destruction as a function of time during this time period?

Nd isotopic data on the oldest (~ 3.8 AE) terrestrial rocks from Isua, West Greenland indicate that the mantle at this time had already suffered substantial depletion in incompatible elements due to earlier continent forming events. These data suggest that the continents started to grow at least 4.2 AE ago.

Isotopic data on young volcanic rocks derived from the depleted mantle show no evidence of this early history. This is most likely due to the fact that recycling of continental material back into the mantle over the history of the earth has erased the evidence.

To quantify this further it is necessary to consider the observed isotopic patterns of Nd, Sr, Hf and Pb through time together with the presently observed age spectrum of crustal rocks. These patterns can be modelled by a transport model in which the continental growth and destruction rates are allowed to vary as a function of time. The results suggest that the mass of the continents at 3.8 AE ago was about 25% of the current continental mass. However, due to the very high recycling rates obtained in the early Archean only a few percent of this crust has been preserved up to the present. Although these results are somewhat model dependent, they could be put on a much firmer basis if a more substantial data base on the isotopic patterns in early crustal rocks were available.

ANORTHOSITIC OCEANIC CRUST IN THE ARCHEAN EARTH. E. Jagoutz*, J.B. Dawson**, S. Hoernes***, B. Spettel*, and H. Wanke*. *Max-Planck-Institut f. Chemie, 65 Mainz, F.R.Germany; **University of Sheffield, Department of Geology, Sheffield S1 3JD, Great Britain; *** Universität Bonn, Mineralog.-Petrolog. Institut, 53 Bonn, F.R.Germany.

Ultra-pure minerals separated from eclogite inclusions in kimberlites (BD 1191, BD 1175, BD 3699 from Roberts Victor Mine, BD 1934 Viesuri Mine, Tanzania) have been analysed for Sm, Nd, Sr, and oxygen isotopes as well as for major and trace elements. Clinopyroxene (cpx) and garnet (gnt) are the only primary mineral phases in these rocks, as confirmed by microscope inspection and isotope and trace element measurements on bulk, mineral phases and their alteration products. The results of the clean mineral phases are given in the table. WR_{calc} is the reconstructed bulk composition excluding all the contamination influences (host magma, post-eruption alterations). For BD 1934, the least altered rock, the whole rock analysis is given for comparison.

From our own and literature data on their chemistry we can distinguish two groups of eclogites:

Type A (Noritic-anorthositic eclogites): LREE enriched; $\delta^{18}O > +6\text{‰}$; HREE depleted (down to $2 \times C1$); commonly positive Eu-anomaly, poikilitic type of texture and cpx exsolving opx lamellae; $Ca > Fe$; (BD 1175, BD 3699).

Type B (Ti-ferrogabbroic eclogites): strongly LREE depleted; $\delta^{18}O < +6\text{‰}$; HREE $> 10 \times C1$; trace elements strongly partitioned into gnt; low $^{87}Sr/^{86}Sr$; very high $^{143}Nd/^{144}Nd$; rutile exsolution in cpx and gnt; $Ca < Fe$; interlocking metamorphic fabric; (BD 1191, BD 1934).

GENETIC RELATIONSHIP:

The oxygen isotopes are primary mantle-derived features of these rocks and are not caused by post-eruption processes, as they were measured on unaltered, clean mineral separates and show a correlation with REE pattern and Sr and Nd isotopes. Garlick et al. (1) suggested a fractionation of oxygen isotopes at mantle P-T conditions, but Clayton et al. (2) demonstrated experimentally and theoretically that this is highly unlikely. Thus we think that the variation of the oxygen isotopes are caused by crustal-level fluid-rock interaction at relatively low temperature.

Muehlenbachs and Clayton (3) have shown that oxygen isotopes variation in MORB basalts caused by the hydrothermal system are in the same range as the observed oxygen isotope variation in eclogites. Low-temperature alteration of basalts by seawater increases their $\delta^{18}O$ value by $1-3\text{‰}$, while high-temperature alteration by post-solidus exchange with seawater causes a decrease of $\delta^{18}O$ by $1-3\text{‰}$.

Type A eclogites cannot be derived from normal MORB-type basalts, because of their low HREE ($2.0 \times C1$) content and positive Eu-anomaly.

We propose the following model to explain the new set of data:

We start with a magma at a shallow level within a subductible segment of lithosphere. This magma differentiates into a mafic and felsic portion. During cooling the hydrothermal alteration system affects the oxygen isotopes as in the present day MORB environment. On subduction this segment is dehydrated and loses its incompatible elements. The more mafic portion, converted to amphibolite facies with light oxygen (BD 1191), will be affected more strongly by dehydration than the less metamorphosed felsic portion (BD 1175).

AGE RELATIONSHIP OF ECLOGITES:

In all cases except BD 3699, minerals are in $^{143}Nd/^{144}Nd$ isotopic disequilibrium at the time of eruption. The cooling temperature for Nd in the gnt-cpx system is uncertain, but there is evidence that the minerals in gnt-lherzolites as well as sp-lherzolites (4) are in isotopic equilibrium at the time of eruption. Thus we think that some of these eclogites might be emplaced into the upper lithosphere or lower crust at the time corresponding to their internal isochron age. This age of BD 3699 is close to the emplacement age of Roberts Victor kimberlite (0.127 b.y.; ref. 5). The calculated WR composition was used to estimate model ages for these rocks (see Table). The data points of WR_{calc} lie on an "isochron" with an age of 2.7 ± 0.1 b.y. and an initial $^{143}Nd/^{144}Nd$ of 0.5096 ± 3 ($\epsilon_0 = +9 \pm 6$). If the $^{87}Sr/^{86}Sr$ of the cpx of BD 1191 (0.70092) is close to the initial ratio of the system, an origin from a depleted reservoir like that of modern MORB is inferred. To clarify this, more isotope measurements on eclogites are needed.

Jagoutz E. et al.

The existence of Archean mantle eclogites is firmly established by our new data. Hence it becomes clear that Archean subduction can be driven by eclogite transformation in the downgoing slab, as it is assumed for the present tectonic environment. Evidence for an anorthositic-noritic-ferrogabbroic oceanic crust in the Archean (6) can be clearly traced until the first ophiolite complexes are appearing (1.7 - 1.5 b.y.). This time might be the onset of today's MORB-production. Relicts of the old type differentiation sequence in the MORB-regime can still be recognized (7).

GEOCHEMICAL CONSTRAINTS:

Type A eclogites have K/Ca ratios down to 300. This is a factor of 10 lower than the c. 1 ratio of 2720. Hofmann and White (8) found that Ca is depleted relative to Rb in MORB and ocean island basalts. There is no doubt that today even the most primitive mantle reservoirs observed are extremely low in Ca. The primitive spinel-lherzolite SC 1 of Jagoutz et al. (9) contained only 1.4 ppb Ca (10). The crustal contribution to the primitive mantle is significant but nevertheless amounts only to 7.7 ppb Ca (11). This adds up to a total of 9.1 ppb Ca, compared to the 85 ppb Ca which one would expect for the primitive mantle assuming a c. 1 K/Ca ratio. If we assume that the missing 76 ppb Ca is locked up in type A eclogites (represented by BD 1175, which contains 8.8 ppm Ca), we calculate that 0.86 % type A eclogites in the earth's mantle is sufficient to account for the observed Ca depletion.

Sample	Sm ppm	Nd ppm	$^{143}\text{Nd}/^{144}\text{Nd}^{**}$	$\delta^{18}\text{O}^*$	$^{87}\text{Sr}/^{86}\text{Sr}$	Intern. Age b.y.	Model Age b.y.
BD 1934 gnt	1.391	0.819	0.52317 \pm 2	5.8			
						1.7	
BD 1934 cpx	1.178	3.032	0.51424 \pm 3	5.7	0.70109 \pm 5		
BD 1934 WR meas.	1.257	2.706	0.51506 \pm 2				
BD 1934 WR calc.	1.264	2.130	0.51563	5.7			3.75
BD 1191 gnt***	0.585	0.114	0.54769 \pm 4	3.8			
						0.90	
BD 1191 cpx	0.255	0.211	0.53358 \pm 4	4.4	0.70091 \pm 3		
BD 1191 WR calc.	0.419	0.162	0.53853	4.1			3.06
BD 1175 gnt ⁺	0.830	1.438	0.51193 \pm 4	6.9			
						0.22	
BD 1175 cpx ⁺	0.756	6.931	0.51152 \pm 6	7.5	0.70155 \pm 5		
BD 1175 WR calc.	0.784	4.836	0.51142	7.3			
BD 3699 gnt	2.810	3.586	0.51554 \pm 2	6.9			
						0.12	
BD 3699 cpx	0.642	2.748	0.51530 \pm 2	7.5	0.70335 \pm 3		
BD 3699 WR calc.	1.697	3.156	0.51544	7.2			4.05

* $\delta^{18}\text{O}$ relative to SMOW in ‰, error 0.1 ‰; ⁺ Sm and Nd spiked on different aliquots; ** normalized to $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$; *** blank correction (2 %) where significant.

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A MODEL FOR CORE FORMATION IN THE EARLY EARTH. John H. Jones and Michael J. Drake, Lunar and Planetary Laboratory, University of Arizona, Tucson, AZ 85721.

The high concentrations of siderophile and chalcophile elements in the Earth's upper mantle have historically presented problems for those geologists interested in the mechanisms and processes of core formation. The most concise statement of the problem was that given by Ringwood [1] via his assertion that the core and upper mantle could never have been in equilibrium, because siderophile elements such as Ni and Ir are several orders of magnitude too abundant in mantle rocks for the mantle to ever have been in equilibrium with iron-nickel metal. Thus, it has been clear for nearly two decades that the magnitude of the siderophile element problem lies not within factors of 2-3 but in factors of 10^2 - 10^4 . Solutions to this problem have been hampered by both a lack of data--experimental data on the chemistry of siderophile elements and analytical data on the abundances of siderophile elements in the upper mantle--and a lack of quantitative models of core formation. Because, while it is true that siderophile elements in the mantle are greatly enriched over concentrations expected in the simplest core formation model, siderophile elements are depleted relative to CI chondrites by factors of ≈ 5 -100. Other than simple disequilibrium between core and mantle [1], the traditional solution to the problem of apparently overly abundant mantle siderophile elements is to mix chondritic material into the upper mantle after core formation. We refer to this class of solutions as "exogenous".

Two basic types of exogenous models have been proposed to account for siderophile and chalcophile element abundances in the Earth's upper mantle. The first model requires that the Earth be depleted in volatiles [2] and that, after a core formation event which extracted the most siderophile elements into the core, additional noble siderophile elements (Pt, Ir, Au) were added as a "late veneer" and mixed into the mantle [3]. The second model postulates a reduced Earth with approximately CI elemental abundances in which a primary core-forming event depleted all siderophile elements in the mantle; 5-10% of more oxidized chondritic material was then mixed into the mantle and a secondary core-forming event depleted only the most siderophile elements; finally, approximately 1% chondritic material was mixed into the mantle to account for the abundances of the noble siderophile elements [4].

We question the plausibility of models which require fine-scale mixing of chondritic material into the upper mantle. The similar concentrations of siderophile and chalcophile elements in lherzolite nodules distributed worldwide argue that such mixing must result in homogeneity at approximately the 1 m scale [3]. Solid state convection/diffusion seems incapable of such fine-scale mixing [5]. Mixing in liquids is more efficient, but large degrees of silicate partial melting will facilitate the separation of magma (into a magma ocean?) from residual solids. Since we apparently do not sample the products of such a magma today, mixing in the liquid state does not resolve the problem. Alternatively, a megaregolith on the early Earth of several kilometers' depth may have been formed. This megaregolith may have been well-gardened to 10-100 m (compare to the 5-10 m lunar regolith), but gardening as deep as 100-700 km seems rather improbable. Finally, any external events affecting the upper mantle of the Earth should also be evident in the Moon (which corresponds to 7% of the volume of the upper mantle); but siderophile and chalcophile element abundance patterns inferred for the mantles of the Earth and Moon differ [6,7]. Thus, there appear to be significant physical difficulties associated with "chondritic veneer" models.

Jones, J.H. and Drake M. J.

These difficulties have led us to investigate an "endogenous" model in which four distinct phases--solid metal, S-bearing metallic liquid, solid silicate and liquid silicate--coexist in chemical equilibrium throughout planet growth. Core formation occurs during planet growth [8], leaving a small residue of solid and liquid metal in the upper mantle (i.e., core formation is not quite 100% efficient). Figure 1 gives calculated whole-Earth abundances of W, Re, Ir, Mo, Ni, Au, P, Ga, Ag, S and Pb (normalized to CI) for a variety of plausible conditions for core formation. These whole-Earth abundances are calculated using experimentally determined partition coefficients, partition coefficients deduced from naturally occurring basalts, and estimates of mantle abundances obtained from measured siderophile and chalcophile element concentrations in mantle materials (Table 1). It should be noted that the estimated abundance of Mo in the mantle is uncertain to a factor of +50%. For all other elements the uncertainties are larger [e.g., 3]. The relative fractions of the four phases and the amount of solid and liquid metal phases remaining in the upper mantle are allowed to vary. With the exception of P and Ni, upper mantle abundances of siderophile and chalcophile elements less volatile than S can be explained by whole-Earth abundances which are chondritic relative and absolute within a factor of two uncertainty if approximately 0.1% solid metal and approximately 1% metallic liquid were trapped by the mantle during core formation. The best fit interpolation among the models is 10%-20% partial melting of the silicates and 2 wt.% - 8 wt.% S in the core. Phosphorus, the element fitted most poorly by this model, is still only a factor of 3-4 lower than elements of similar volatility (Au, Ga). Given the inherent uncertainties in any model of core formation, it is not clear that the discrepancy is significant. Thus, with simple assumptions, reasonable whole-Earth compositions and geologically plausible conditions, most siderophile and chalcophile element abundances in the mantle may be explained by a process indigenous to the Earth, obviating the need for although not excluding subsequent mixing of late-stage "chondritic veneer" material.

Metal is not stable in the mantle today. Thus, for our model to be valid, the upper mantle must have become more oxidizing over geologic time--a trend consistent with the subduction of oxidized materials into the mantle. A major consequence of our model is that fractionation of U from Pb occurred both prior to Earth assembly as well as during core formation, since no model allows Pb in the whole-Earth to be within a factor of four of chondritic. Our best-fit model is consistent with S being an abundant light element in the Earth's core, but does not exclude other elements as significant constituents.

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Jones, J.H. and Drake, M.J.

ORIGINAL PAGE IS
OF POOR QUALITYTABLE 1: PARTITION COEFFICIENTS (D'S)
AND INFERRED MANTLE SIDEROPHILE ABUNDANCES*

	W	Re	Ir	Mo	Ni	Au	P	Ga	Ag	Pb
D ₁	1	5×10^{-4}	5×10^{-5}	0.001**	0.003**	10^{-4}	0.24	0.8	0.01	0.15
***D ₂	0.01	~1	~500	0.01	10	~1	0.02	0.4	0.4	0.4
D ₃	35	85	85	2.45	1.3	1.3	1.7	6	0.01	0
C _M	0.05 ^[4]	0.0075 ^[3]	0.0075 ^[3]	0.02 ^[4]	0.20 ^[13]	0.02 ^[3]	0.025 ^[14]	0.28 ^[9]	0.11 ^[3]	0.09 ^[10]

D₁ = silicate liquid/metallic liquid; D₂ = silicate solid/silicate liquid
 D₃ = metallic solid/metallic liquid; C_M = mantle concentration/CI

T = 1250-1270°C; -log f_{O₂} = 12.5-13.0; metallic liquid contains ~25 wt.% S.

*W, P and Mo abundances from W-La, P-La and Mo-La correlations in naturally occurring basalts; all other abundances from mantle isherzolite analyses.

**D₁(Mo) estimated from our data and that of [11]; D₁(Ni) estimated from our data and that of [12].

***D₂'s estimated from naturally occurring basalts; all other D's from our laboratory partitioning studies.

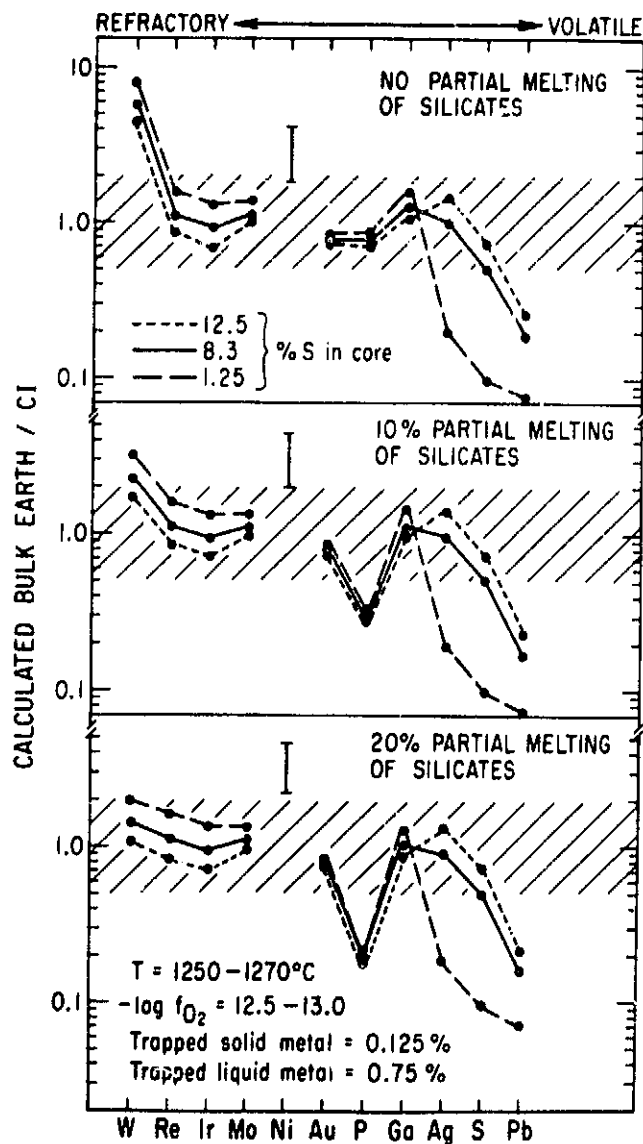


Figure 1. Results of endogenous core formation models. Calculated whole-Earth concentrations (relative to CI chondrites) are shown for various solid metal/liquid metal ratios and various degrees of silicate partial melting. Calculations for Ni are shown as a bar--denoting uncertainty in the appropriate D₁ for Ni. The solid metal/liquid metal ratio is presented as wt.% S in the core; S in the metallic liquid is taken to be 25 wt.% and S in the solid metal is negligible. The shaded regions are factor-of-two envelopes about CI abundances which reflect the magnitude of the uncertainty in mantle siderophile element abundances. Preferred models permitting approximately chondritic relative abundances in the whole-Earth for elements less volatile than S require that ~0.1% metal and ~1% metallic liquid be trapped by the mantle during core formation and that the silicates be 10-20% partially melted.

IMPLICATIONS FOR THE EARTH OF THE EARLY DYNAMICAL ENVIRONMENT. W. M. Kaula and S. A. Cooperman, Dept. of Earth & Space Sciences, University of California, Los Angeles 90024.

The formation of the Earth, which took 100 million years or less about 4.50 billion years ago, was mainly from sizeable bodies: perhaps moon-sized. The principal reason for believing this is that any mechanism ever proposed for making planetesimals depends on conditions which will lead to the formation of many planetesimals. No one has ever conjectured a "magic suppressor" which allows the formation of four or five kernels for the terrestrial bodies but prevents formation of any other similar bodies. Models of interaction among small planetesimals which take into account only close encounters all lead to the formation of moon-sized objects (1,2), thus leading to several 100 in the inner solar system. Longer-term interactions, such as secular resonance sweepings (3), are needed to get these planetesimals together to form the observed terrestrial bodies.

The consequence is that the formation process is dominated by great impacts, most of whose energy is retained as heat at time of impact (4,5,6). What might be called the ten-little-indians scenario seems the most plausible working hypothesis: before there were four terrestrial planets there were five; before there were five, there were six; and so forth. Hence the terminal stages of planet formation were probably dominated by great collisions, involving Mars-sized bodies. Evidences that this small-number statistics situation prevailed in the early solar system are the great differences among the planets: between Venus and the Earth, in rotation rate and presence of a satellite; among Venus, Earth, and Mars, in primordial argon retention; between Mercury and the rest, in iron content; and among all four, in size.

The energy released in a collision between a Mars-sized and an Earth-sized planet is more than 10^{38} ergs, about 10^6 times that creating Mare Imbrium and comparable to the total radioactive heat of the Earth throughout its history. While considerable model definition and calculation is needed, it probably is enough to put two percent of the mass involved into orbit in a highly devolatilized state. It also would have considerable implication for the Earth's state, melting or vaporizing an appreciable fraction and stimulating core formation and outgassing. How efficient the outgassing is problematical. The two orders-of-magnitude difference in primordial argon retention between Venus and the Earth suggests that it was very efficient (7), and further that there was little subsequent volatile acquisition. Most of the energy pulse from such a collision would be removed by enhanced convection within a few million years, a process which would be

Kaula, W. M.

more conducive to formation of a primitive crust than was the collision event itself.

The dynamical settling down of the inner solar system was probably much more prolonged than compatible with an exponential fit to the tail of the classical Safronov sigmoidal growth curve (8). Firstly, the time scale of clearing out all bits and pieces of the inner solar system-- out to, say, 3.28 AU, the Jupiter 1:2 resonance location-- would have been a sizeable multiple of perihelion revolution periods (10^{**5} years), rather than of orbital years. Secondly, the densities of Uranus & Neptune indicate they formed from icy bodies. If this was by the same mechanism as the formation of the terrestrial planets from rocky bodies, then the time scale would have been more than ten times as long, and considerable matter would have been scattered to the inner solar system. A counter-indicator to this process is the great difference in primordial gas abundances of Venus and the Earth. While impactors from the outer solar system would come at such high velocities as to lead to net mass decreases, they would be messy enough events that it is difficult to imagine their not contributing some volatiles to the planets. In addition, the exceptionally small eccentricity of Neptune's orbit and the deep resonance of Pluto's orbit, suggest resonance sweepings, related to loss of appreciable mass from the solar system, were important, greatly shortening the time scale of formation.

The main import for early Earth evolution is that impacts continued to be important shapers of the primordial crust for a few 100 million years without there necessarily being a "late heavy bombardment" about 4.0 b.y. ago. The resetting of radioactive clocks on the Moon which may (or may not) be needed could have accomplished by few events-- one or two or three. The sort of event which recurred every few m.y. or 10 m.y. on Earth in its first three or four 100 m.y. would have been a super-Imbrium: 10^{**33} or 10^{**34} ergs, enough to punch through the crust but not to perturb the core. The sum of these events would also be enough to reset all radioactive clocks and may have inhibited craton stabilization.

The influence of impacts on craton stabilization is problematical. While the super-Imbrium impacts made great surficial splashes, they were a minor source of energy for differentiation and tectonism by 100 million years after formation: i.e., since 4.4 b.y. ago. More important were primordial heat (core separation as well as accretion) and radioactivity, the sum of which would have been at least 10^{**35} ergs / m.y. 4.3 b.y. ago (compared to 10^{**34} ergs / m.y. now). Certainly the great impacts promoted lateral heterogeneity, and may have influenced the pattern of mantle convection by removal of the sialic layer over appreciable extents, leading to rifting of cratons.

Kaula, W. M.

The central problem of craton stabilization is, of course, the development of lateral heterogeneity. It may have been a rather Darwinian chance process: sizeable pieces of sialic crust which blundered over regions of upper mantle of high Mg: Fe ratio and depleted in radioactive lithophiles would be the ones to survive. A question is how much mantle convection tended to homogenize the mantle. The lithosphere would have been too thin for Phanerozoic style subduction. Anorthosite or basalt differentiated from a convective upstream would have tended more strongly to remain at the surface. The steeper temperature gradients would have led to spreading out of sialic accumulations, producing a thin crust under a global ocean (9). Phase transitions can be invoked to recycle sial below accumulations of more than 30 km, but this does not explain why the mean crustal thickness remaining is much less.

Geosynclines plausibly could have evolved without subduction. There were certainly convective downstreams, ten times as vigorous as now, under the crust. These would have created appreciable depressions in the surface which would have been accumulation sites for sediments which further depressed the surface until the lowermost sial was deep enough for the phase transition, which would have accelerated the process further. This would have occurred on a much shorter time scale than in Phanerozoic, so that the heating by radioactivity and insulation (10) would not have limited the process until a much greater geosyncline had developed. Another factor would have been that the more primitive differentiates of the early crust would not have had as much radioactives (though having higher generation rates) as Phanerozoic continental crust. In this manner, there could have been appreciable recycling of sial on the early Earth, the sixfold difference in gravity leading to a very different regime from that on the early Moon.

After the accumulation of the Earth, during which core formation certainly occurred, further impacts probably influenced the locations of rifting centers in the system of mantle convection and crustal differentiation. They may have affected craton stabilization by promoting lateral heterogeneity, but had little influence on the key problem of early recycling of sial.

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A REVIEW OF TECTONIC ASPECTS OF THE LIMPOPO BELT AND OTHER ARCHEAN HIGH-GRADE GNEISSIC TERRANES. W.S.F. Kidd, Dept. of Geological Sciences, S.U.N.Y. at Albany, Albany, NY 12222.

Published information on the Archean high-grade terrains varies a great deal in the detail available. Such information as exists indicates marked differences in the lithic types and proportions present in the central Limpopo belt compared with the better studied of the other Archean high-grade terrains. These differences may be important because they are expressed by the presence in the Limpopo belt of subordinate, but significant quantities (about 5% each) of two rock suites likely to have formed on a shallow marine platform of significant size (Eriksson and Kidd, in prep.). These suites consist of thick sections dominantly consisting of either carbonate and calc-silicate, or of pure meta-quartzites, often fuchsite-bearing, whose lithic characters are unlike those expected for metacherts but are very like those expected for platform arenites. Isotopic ages suggest these sediments are probably older than 3.3 Ga and younger than 3.5 Ga. They are similar (except in metamorphic condition) to preserved sediments deposited in later geologic times on the interior and on the Atlantic-type margins of equatorially-located continents, for example in the Coronation belt of early Proterozoic age at the western edge of the Slave Province in Canada.

There is a common overall lithic/structural sequence in the two largest and locally well-described Archean gneissic terranes (West-Greenland/Labrador and central Limpopo belt). This consists of an older granitoid gneissic basement, cut by a mafic dike swarm, and overlain by a cover sequence of largely metavolcanic and metasedimentary rocks, some aspects of which are discussed above. Both basement and cover are involved in severe deformation and metamorphism to granulite facies, which is proposed to have involved large nappes early in the deformation and subsequent multiple refolding of those structures. Some syn- to late-kinematic granitoid plutons intrude the structural assemblage, presumably emplaced during the earlier stages of uplift and cooling of the rocks now exposed. Peak metamorphic assemblages indicate a maximum depth of burial of about 30 km which implies a double crustal thickness at that time. Such a sequence of events characterizes younger terrains of the Proterozoic and Phanerozoic where there has been rifting of continental lithosphere, followed by deposition of volcanic and sedimentary material at an Atlantic-type margin, followed by convergence and collision with another continental object. The structural and metamorphic condition of the rocks largely reflect the final collision and accommodation to it by crustal thickening. Record of the pre-collisional convergent phase of the Wilson cycle in terms of sedimentary, volcanic or plutonic lithologies of accretionary or arc environments may well be contained in these Archean terrains but is harder than the other items to recognize clearly.

A further aspect which could help tectonic understanding of these gneissic terranes is their large-scale relationship to lower-grade Archean rocks. There is only one example (the Limpopo belt) that is relatively well-mapped where it is possible to observe both margins of a large Archean gneissic terrane in their Archean relationship to adjacent terranes. The metasedimentary component of the central Limpopo is readily distinguished from the greenstone-granodiorite assemblage of the adjacent Kaapvaal and Zimbabwe cratons, although the metamorphic and structural effects of the Limpopo belt extend a significant distance into those cratons. The southern margin of the central Limpopo belt is a wide (20 km) zone of vertically-dipping, horizontally-lineated mylonites, clearly representing the deeper ductile levels of a major strike-slip fault, probably having dextral offset. The northern margin is a similar but southward-

dipping zone of ductile shear, proposed to have major thrust displacement, although a subhorizontal fabric lineation suggests significant strike-slip motion as well (also dextral). Other proposed northwest-directed thrusts occur in the Zimbabwe craton to the north, in the zone in which the Limpopo metamorphism declines. In a first-order sense, the cross-section derived from these structural observations, and the metamorphic data (taken as indicating the depth of material eroded) shows a strong family resemblance to a section across the Himalaya and Tibet, where the thrusts over the Zimbabwe craton are equivalent to the Himalayan thrusts, the metamorphic zone of the central and southern Limpopo belt to the overthickened crust of the High Himalaya and the Tibetan Plateau, and the large strike-slip fault separating the central and south portions of the Limpopo belt equivalent to the KunLun or Kang Ting Faults that allow portions of Tibet to attempt to escape the collisional zone. Available isotopic data indicate that collisional tectonics, uplift, erosion and cooling in the Limpopo belt occurred between about 3.0 and 2.6 Ga ago.

The three aspects discussed lead to the conclusions that a) continental fragments large enough to provide a substrate for significant platform arenite and carbonate sedimentation existed by 3.3 - 3.5 Ga ago; b) Wilson-cycle tectonics seems to adequately explain most major features of the Archean gneissic terranes; c) Tibetan-Himalayan style collisional tectonics 2.6 Ga and older accounts for the large-scale relationships between the Limpopo belt and the adjacent Archean greenstone-gneissoid terrane cratons. By inference, other more fragmentary Archean gneissic terranes may have once been part of such collisional zones.

A Review of Noble Gas Geochemistry in Relation to Early Earth History

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One of the most fundamental noble gas constraints on early earth history is derived from isotopic differences in $^{129}\text{Xe}/^{130}\text{Xe}$ between various terrestrial materials. The short half-life (17 m.y.) of extinct ^{129}I , parent of ^{129}Xe , means that these differences must have been produced within the first 100 m.y. after terrestrial accretion. The identification of large anomalies in $^{129}\text{Xe}/^{130}\text{Xe}$ in mid-ocean ridge basalts (MORB), with respect to atmospheric xenon, suggests that the atmosphere and upper mantle have remained separate since that time (1-2). This alone is a very strong argument for early catastrophic degassing, which would be consistent with an early fractionation resulting in core formation. However, noble gas isotopic systematics of oceanic basalts show that the mantle cannot necessarily be regarded as a homogeneous system, since there are significant variations in $^3\text{He}/^4\text{He}$, $^{40}\text{Ar}/^{36}\text{Ar}$, and $^{129}\text{Xe}/^{130}\text{Xe}$ (2-3). Therefore, the early degassing cannot be considered to have acted on the whole mantle.

The specific processes by which degassing presently takes place can also be used to decipher the record of early degassing. At present, gases are brought to the surface, through the crust, by magma transport and exsolution of a volatile phase. In the presence of a very thin crust, or a magma ocean, the degassing would be more directly related to equilibrium between magma and the early atmosphere. Thus, the specific mechanisms of degassing, in particular the thickness and growth of the early crust, is an important variable in understanding present-day noble gas inventories.

Another constraint can be obtained from rocks that are thought to be derived from near the lithosphere-asthenosphere boundary: ultramafic xenoliths. Preliminary results of a helium isotopic study of spinel lherzolites from various localities show that the $^3\text{He}/^4\text{He}$ ratios of the xenoliths from sub-continental areas are significantly lower than those from oceanic regions. $^3\text{He}/^4\text{He}$ ratios from San Carlos, Kilbourne Hole, and Massif Central are 5-7x the atmospheric ratio, while lherzolite from Salt Lake Crater in Hawaii is closer to the MORB value of 8.4x atmospheric. This may indicate that the lithosphere once formed in oceanic regions is presently stored beneath the continents. This simple interpretation is complicated by the small data set, the complex deformation history that the xenoliths have undergone, and by the fact that much of the helium in these samples is contained in fluid inclusions (as determined by crushing in vacuo). Therefore, the isotopic variability may also reflect more recent volatile transport processes.

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ARCHEAN SEDIMENTARY SYSTEMS AND CRUSTAL EVOLUTION, Donald R. Lowe
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Our current knowledge of preserved Archean sedimentary rocks suggests that they accumulated in at least three major depositional settings. These are represented generally by sedimentary units (I) in early Archean, pre-3.0 Ga-old greenstone belts, (II) on late Archean sialic cratons, and (III) in late Archean, post-3.0 Ga-old greenstone belts.

(I) The principal volcanic sequences of most early Archean greenstone belts, including the Barberton greenstone belt, South Africa, and belts in the eastern Pilbara Block, Western Australia, contain assemblages of sedimentary units deposited on low-relief, anorogenic but rapidly subsiding, shallow-water platforms constructed largely of mafic and ultramafic volcanic rocks (1). The main sediment types included freshly erupted or only slightly re-worked pyroclastic debris; orthochemical precipitates including carbonate, evaporites, and silica; and biogenic deposits now represented by black carbonaceous chert, banded chert, and stromatolitic units (1, 2, 3). Terrigenous deposits, such as sandstone and shale derived by weathering and erosion of older rocks, are rare. Algoma-type iron formation, volcanogenic massive sulfide deposits, and turbidites, all of which usually form under deeper, fully subaqueous conditions, are also poorly developed. Detrital modes of clastic units are dominated by volcanic and intraformational sedimentary components and essentially lack coarse α -quartz, potash feldspar, and metamorphic and plutonic rock fragments. These sedimentary rocks reflect sedimentation on large, stable, shallow-water, simatic platforms that were far removed from sources of sialic debris.

The upper, sedimentary parts of these early Archean greenstone belt sequences are made up largely of terrigenous detrital units reflecting orogenesis and deep erosion (4). The debris includes material eroded from both the underlying greenstone belt sequence and quartzose intrusive and metamorphic rocks. Deposition took place in a complex association of environments ranging from alluvial to deep-sea (4). These units mark the final stage of greenstone belt evolution and the regional transition from simatic platforms to sialic continents.

(II) Following early Archean cratonization, the South African Kaapvaal and Australian Pilbara cratons were subject to deep erosion and, in late Archean time, were the sites of deposition of thick sedimentary and volcanic sequences. These include the Fortescue Group on the Pilbara Block and the Pongola and Witwatersrand Supergroups and succeeding units in South Africa. Although volcanism was widespread during the accumulation of many of these sequences, the sedimentary rocks clearly reflect deposition on relatively stable blocks of continental crust. They directly overlie the eroded plutonic and greenstone basement of early Archean age, are much less deformed and metamorphosed than underlying greenstone-belt rocks, have not been extensively intruded by granitoid plutonic rocks, and contain large amounts of continent-derived arkosic and quartzitic debris.

(III) Later Archean, post-3.0 Ga-old greenstone belts in Canada and much of the Yilgarn Block, Western Australia, include sedimentary sequences that differ significantly from those in the older belts (1). Coarse terrigenous clastic units occur throughout the volcanic section instead of just at the top, as in the older belts. They are made up of conglomerate, sandstone, and shale derived largely by erosion of the greenstone belt sequence itself. A small amount of detritus from plutonic sources is present throughout the sections, however, including material eroded from both penecontemporaneous intrusive units and older sialic basement. Deposition was mainly by sub-

Lowe, D. R.

aqueous sediment gravity flows, including turbidity currents and debris flows. Shallow-water deposits in the volcanic sections developed principally around high-standing felsic volcanic vents. Carbonaceous cherts, banded cherts, and evaporites are rare, whereas volcanogenic sulfide deposits and Algoma-type iron formation are widespread. Deposition took place under subaqueous, in large-part deep-water, tectonically unstable conditions (1).

As in the older belts, the uppermost clastic units at the top of the greenstone belt sequence reflect orogenesis. They were deposited in an association of environments ranging from alluvial to deep-sea (5).

The sedimentology and comparative sedimentology of these Archean sedimentary systems suggest a number of major inferences about the character and evolution of the Archean crust. (A) The well-preserved early Archean greenstone belts formed as large, high-standing, anorogenic simatic blocks that were not underlain by or adjacent to significantly older continental crust. They underwent cratonization which age relationships of associated plutonic rocks suggest proceeded more-or-less concurrently with deposition of the greenstone belt sequences (6, 7).

(B) The late Archean cratons, formed by cratonization of the early Archean greenstone belts, served as sites for the accumulation of late Archean volcanic and craton-derived, shallow-water shelf sequences. These sequences are generally similar to cratonic sediments that have accumulated throughout later geologic time and are clearly distinct from greenstone belt sediments.

(C) The late Archean greenstone belts also represent large, probably high-standing but largely subaqueous and tectonically active depositional sites (1). They were closely associated with significantly older sialic blocks, but evidently these blocks were not of sufficient extent or thickness to form stable, durable cratons. They may represent incompletely cratonized areas or microcontinents. These belts also underwent cratonization more-or-less concurrently with accumulation of the greenstone belt sequence to form stable continental blocks that hosted the deposition of early Proterozoic shelf sediments. The co-existence in the late Archean of large, stable cratons, in South Africa and Western Australia, and the younger greenstone belts emphasizes the sedimentological and tectonic contrasts between these terranes and argues strongly against suggestions that these and, by analogy, other greenstone belts developed upon thick continental crust.

These results suggest that the Archean was characterized by at least two distinctive and largely diachronous styles of crustal evolution. Thick, stable early Archean simatic platforms, perhaps analogous to modern oceanic islands formed over hot spots, underwent a single cycle of cratonization to form stable continental blocks in the early Archean. Later-formed Archean continents show a two-stage evolution. The initial stage is reflected in the existence of older sialic material, perhaps representing incompletely cratonized areas or microcontinents of as yet unknown origin. During the second stage, late Archean greenstone belts, perhaps analogous to modern magmatic arcs or back arc basins, developed upon or adjacent to these older sialic blocks. The formation of this generation of Archean continents was largely complete by the end of the Archean. These results suggest that Archean greenstone belts may represent a considerable range of sedimentological and tectonic settings.

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THE ORIGIN OF CONTINENTAL CRUST: OUTLINES OF A GENERAL THEORY

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The origin of Earth's continental crust remains one of the most difficult problems in geology. However, several lines of evidence are beginning to converge toward the long-sought "general theory" called for by Tuzo Wilson (1949): (1) Comparative planetology, (2) Geologic mapping of shields, and (3) New data on the nature of the lower continental crust.

Comparative Planetology

The main stages in the crustal evolution of silicate bodies Moon-sized and larger include a period of early global differentiation, impact bombardment, and a second differentiation involving generation of basic magma which is still continuing on the Earth as formation of oceanic crust. The "first differentiation" (Lowman, 1976), inferred directly for the Moon and indirectly for Mars and Venus (Solomon, 1980), resulted in production of global crusts of which the lunar highlands are the type example (though probably not representative petrologically). A degree of differentiation can occur even in bodies as small as Vesta (Chapman, 1982) although the inferred carbonaceous chondrite composition for Ceres indicates that it requires special conditions not now understood. The collective evidence from comparative planetology shows clearly that the Earth had the thermal energy and volatile content to undergo a first differentiation that produced the earliest continental crust. Evidence to be cited indicates that, contrary to general opinion, this primordial crust may survive as the lower levels of existing continental crust.

Shield Mapping

Geologic and geophysical studies of shield areas, Canada in particular, are beginning to indicate the general nature of crust-forming processes and their timing. An important characteristic of most shields is that they do not show a systematic concentric pattern of age provinces or fold belts that might indicate growth by lateral accretion, whether by ensimatic eugeosynclines or acquisition of allocthonous terranes. Second, most age provinces except the very oldest are demonstrably underlain by still older rocks, generally high-grade gneisses of intermediate composition, a relationship widely though not universally interpreted to mean that shield evolution has been broadly a process of reworking rather than addition of new material from the mantle (except for greenstone belts). Third, there is considerable indirect evidence, in Archean sediment compositions, for the existence of sialic crust before formation of the oldest exposed rock (e.g., Goodwin, 1981; Henderson, 1981), although some authorities such as Anhaeusser (1981) consider the earliest crust to have been oceanic. Fourth, gneisses of intermediate to acidic composition dominate the Archean provinces areally; i.e., the "granites" of granite-greenstone terranes are the major component of such areas. Finally, the Archean crust appears to have been about as thick by 3.8 b.y. ago as it is now (Condie, 1982). Collectively, these lines of evidence from shield mapping suggest that much of the continental crust formed in the Archean, and can be interpreted as implying the formation of at least some sialic crust in the first few hundred million years of the Earth's history, whether it survives today or not.

Lowman, P.D.

Nature of the Lower Continental Crust

The lower continental crust, formerly very poorly understood, has recently been investigated by various geological and geophysical techniques that are beginning to yield a generally agreed-on though still vague model (Lowman, 1984). As typified by at least some exposed high-grade terranes, such as the Scottish Scourian complex, the lower crust in areas not affected by Phanerozoic orogeny or crustal extension appears to consist of gently-dipping granulite gneisses of intermediate bulk composition, formed from partly or largely supracrustal precursors. This model, to the degree that it is correct, has important implications for early crustal genesis and the origin of continental crust in general. Most important, it implies that except for areas of major overthrusting (which may of course be considerable) normal superposition relations prevail, and that since even the oldest exposed rocks are underlain by tens of kilometers of sial, true primordial crust may still survive in the lower crustal levels (cf. Phinney, 1981).

The inferred supracrustal origin of the pre-metamorphic lower level rocks is debatable; much of the debate centers on the origin of the gray gneisses making up most of the high-grade shield terranes (Windley, 1976). The widespread occurrence of undoubted metasedimentary remnants, though minor in volume, argues for a supracrustal source for at least some of the lower crust (Walker and Dennis, 1983; Lowman, 1984). A combination of sedimentary and volcanic parent rocks seems likely, although it has been shown that the bulk composition of Scourian granulites corresponds to intermediate volcanics.

A General Theory: Major Aspects

Although it would be premature to propose a detailed comprehensive theory for the origin of continental crust, the outlines at least of such a theory can be suggested.

1. Most continental crust formed in the Archean; crustal evolution since then has been largely reworking and recycling (the latter accounting for isotopic evidence widely interpreted to imply addition of new sial from the mantle).
2. Primordial crust, from the first few hundred million years of the Earth's history, still survives near the base of the continental crust in granite-greenstone terranes. It was formed by a combination of poorly-understood igneous and exogenic processes that produced a global crust originally consisting largely of supracrustal rock, now the granulitic gneisses of high-grade Archean terranes.
3. Impacts analogous to those responsible for lunar mare basins broke up the initial crust about 4 billion years ago, initiating mantle upwelling, basaltic volcanism, and an early period of rapid plate tectonics (Frey, 1980). These processes slowed gradually, transitioning to essentially actualistic plate tectonics about 2.5 billion years ago. True continental growth since the Archean has been minor.

The present continents are, in this concept, the greatly-altered remnants of an original global crust, most of which was formed in the early Archean. The concept can be tested by deep drilling in the "granitic" part of a granite-greenstone terrane, such as the Superior Province. Seismic reflection profiling of such areas is also recommended.

ORIGIN OF CONTINENTAL CRUST

Lowman, P.D.

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THE ARAVALLI SEQUENCE OF RAJASTHAN, INDIA: A PRECAMBRIAN
CONTINENTAL MARGIN?

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The extent to which plate tectonics in its present form operated during the Precambrian is unknown, but is a subject of considerable current interest. We describe here a remarkable succession of Precambrian rocks in Rajasthan, Northwestern India, which may help to shed more light on this question. These preliminary observations are garnered mainly from the published literature, but include also some initial geochemical data from a recently initiated study of the area.

The most detailed early study of Rajasthan geology was made by Heron (1), and most subsequent work has followed his subdivisions, with a few modifications. Three major Precambrian units were recognized by Heron: the Banded Gneiss Complex (BGC), which he considered to be the regional basement, and two overlying groups dominated by sedimentary rocks, the Aravalli and Delhi sequences. Most workers have explicitly or by implication suggested that both of the latter are typical "orogenic" sequences (e.g., ref. 2). Major unconformities mark the base of both the Delhi and Aravalli successions.

Very few reliable geochronologic data are available for these rocks. We (ref. 3) have measured a whole-rock Sm-Nd isochron incorporating mafic and felsic components of the BGC which gives an age of 3.5AE. However, this result strictly applies to only a small portion of the BGC near the city of Udaipur; the complex as a whole is very large and complicated and may contain components with a variety of ages. Nevertheless it appears that the basement upon which the Aravalli rocks were deposited is ancient. The stratigraphically youngest of the Precambrian units identified by Heron comprises the Delhi sequence. Choudhary and Gopalan (4) have set lower limits on the time of deposition of Delhi sediments by measuring Rb-Sr whole rock isochrons for a number of granites intrusive into the Delhi rocks. At least some of these granites are believed to be coeval with the first deformation of the Delhi sequence. The Rb-Sr isochrons indicate ages of 1500-1700 my for the granites. Thus the Aravalli rocks, sandwiched between the BGC and the overlying Delhi sequence, have an age between 1700 and 3500 my. Nd model ages for several metavolcanics from the basal part of the Aravalli succession near Udaipur are in the range 2300-2600 my using chondritic parameters, and slightly higher if a depleted mantle is assumed.

Were the Aravalli rocks produced along an active continental margin during the late Archean or early Proterozoic? This question cannot be answered unequivocally with the data at hand, but the evidence is tantalizing as is clear from a brief description of the sequence. The base is frequently marked by a conglomerate and/or quartzite band resting unconformably on the underlying BGC. In some regions, notably the area around Udaipur, a relatively thick section of metavolcanic rocks occurs above the basal quartzite/conglomerate. These comprise both flows with relict amygdules, and tuffs. Major element compositions for these rocks indicate that they are medium to high-K tholeiitic andesites and andesites. Most samples we have analyzed have SiO₂ between 54 and 57% and K₂O between 0.9 and 1.9%.

Maddougall, J.D. et al.

They are distinguished from most modern orogenic andesites by relatively high TiO_2 (typically 1.1 - 1.4%), although this characteristic is not unique (5). The metavolcanic rocks are overlain by a thick succession of metasediments in which both shallow and deep water facies have been recognized by Roy et al. (6). The shallow water facies include greywackes and impure marbles and quartzites, as well as phosphorites. The major component of the Aravalli sediments is described simply as phyllite (1). Minor amounts of ultramafic rocks are also present in the area; these were said by Heron (1) to be intrusive into the metasediments although no clear description of the contacts is given. They follow approximately the boundary between the shallow and deep water facies of Roy et al. (6). The entire section is complexly folded, with at least two superimposed deformations evident, and no reasonable estimate of its original thickness can be made. Taken together, these data show that the Aravalli sequence has a number of characteristics generally ascribed to active continental margins. Although much more work is required to bear this out, the evidence suggests that the processes operating in such an environment in the early Proterozoic or late Archean were not vastly different from today.

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Carbon and nitrogen in the deep Earth

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The existence of carbon in the mantle is well documented and its origin is of especial interest in view of the diversity of its physical and chemical forms (i.e. graphite, diamond, CH_4 , CO , CO_2) and its wide isotopic variability ($\delta^{13}\text{C}$ ranging from -35‰ to $+3\text{‰}$). The isotope geochemistry of nitrogen, however, an element present in volcanic gases and closely associated with carbon in sedimentary organic matter and in diamonds, remains poorly understood. Employing novel high-sensitivity mass spectrometric techniques we have initiated a study of C-N isotope systematics in basic igneous rocks and mantle materials in an attempt to ultimately constrain the primordial isotopic compositions of these elements and their subsequent fate during the early history of the Earth. In this abstract we will summarise preliminary results obtained for a selection of submarine basaltic glasses and diamonds, concentrating on (i) the problem of whether isotopes are significantly fractionated during igneous processes, and on (ii) the effects of crustal recycling.

High resolution stepped heating reveals two isotopically distinct forms of carbon and nitrogen in submarine basalt glasses. Isotopically light carbon ($\delta^{13}\text{C} = -24\text{‰}$ to -30‰) and heavy nitrogen ($\delta^{15}\text{N} = 0\text{‰}$ to $\sim +30\text{‰}$) are released by combustion (or pyrolysis in the case of nitrogen) from 400 °C to 600 °C; isotopically heavy CO_2 ($\delta^{13}\text{C} = -3\text{‰}$ to -16‰ , see below) and light nitrogen ($\delta^{15}\text{N} = -0\text{‰}$ to -36‰) are liberated from vesicles from 600 °C to 1200 °C. It is known that organic material such as kerogen combusts at comparatively low temperatures (200 °C to 500 °C) yielding isotopically light carbon and heavy nitrogen (1). This and other lines of evidence suggest that the low temperature release component may be surficial organic contamination (2).

A survey of MORB and ocean island basalt (OIB) glasses demonstrates that the $\delta^{13}\text{C}$ of carbon released above 600 °C (magmatic carbon) varies from -2.8‰ to -6.7‰ and the $\delta^{15}\text{N}$ of magmatic nitrogen varies in these samples from -20‰ to -36‰ . The bulk isotopic analysis of carbon in basalts usually yields $\delta^{13}\text{C}$ values within the general range of -10‰ to -30‰ . To account for the discrepancy between these carbon values and those believed to be more typical of mantle (e.g. -10‰ to -2‰ , as shown by the majority of diamonds) some workers (3, 4) have proposed that extensive fractionation caused by the loss of up to 97% of the original carbon takes place during uprise and degassing of the magma. Noble gas systematics are not compatible with such a model (5) and the experiments described in this study confirm that the composition of magmatic carbon in MORB and OIB is not dissimilar to that of typical diamonds. A further implication is that bulk measurements of carbon and nitrogen in basalts probably includes, in addition to indigenous C + N, an organic contaminant component. We see no compelling evidence at present that significant fractionation of C + N isotopes takes place during the uprise and eruption of submarine basalts and that the observed isotopic composition of magmatic C + N are not representative of the upper mantle.

Mattey, D.P. et al.

Whilst insufficient $\delta^{13}\text{C}$ and $\delta^{15}\text{N}$ data exists at the time of writing to allow comparisons between MORB and OIB, we do note that back-arc basin basalt (BABB) glasses contain isotopically distinctive carbon and nitrogen. The $\delta^{13}\text{C}$ and $\delta^{15}\text{N}$ values of magmatic C + N in glasses from the Scotia Sea and Mariana Trough range from -7.7‰ to -16.3‰ and -22‰ to 0‰ respectively. Compared to typical MORB and OIB values, these results are on average $\sim 5\text{‰}$ lighter in $\delta^{13}\text{C}$ and at least 10‰ heavier in $\delta^{15}\text{N}$ and can be explained by the mixing in the source regions for BABB magmas of MORB-like CO_2 ($\delta^{13}\text{C} \approx -6\text{‰}$; $\delta^{15}\text{N} \approx -25\text{‰}$) with an organic carbon component from subducted pelagic sediments ($\delta^{13}\text{C} \approx -25\text{‰}$; $\delta^{15}\text{N} \approx +30\text{‰}$). A corollary of such a mixing model is that continentally derived sediment may survive island arc magma extraction and is thus recycled into the mantle, including significant amounts of pelagic carbonates ($\delta^{13}\text{C} \approx 0\text{‰}$).

Whilst the majority of diamonds from southern Africa possess a relatively narrow range of $\delta^{13}\text{C}$ values with a mode around -6‰ (6), diamonds exhibit a total range in composition of close to 40‰ (e.g. 7). Internal variation of $\delta^{13}\text{C}$ within single diamonds can be complex but is generally restricted to $< 5\text{‰}$. As more data becomes available from other areas it is clear that the isotopic composition of carbon in the upper mantle is grossly heterogeneous. Our measurements of nitrogen abundance and isotopic composition in Type I diamonds show large ranges in both abundance (30 ppm to 1250 ppm) and isotopic composition ($\delta^{15}\text{N} = -13\text{‰}$ to -37‰). However, nitrogen abundance and composition varies within individual diamonds by up to a factor of 2. Although the cause of this considerable internal isotopic heterogeneity is not understood at present, viewed as a whole the nitrogen isotopic composition of diamonds is comparable to the composition of magmatic nitrogen in MORB and OIB.

A compilation of available $\delta^{13}\text{C}$ and $\delta^{15}\text{N}$ measurements on basaltic glasses and diamonds is presented in the Figure. Also plotted for comparison are the compositions of marine carbonates, atmosphere, plants and oils. The average isotopic compositions of ('juvenile') C and N in the upper mantle, as measured in MORB, OIB and diamonds are around -6‰ and -25‰ , respectively. We suggest that recycling of pelagic sediments re-introduces relatively light carbon and relatively heavy nitrogen into the mantle, shifting the composition of BABB and some diamonds away from 'juvenile' values towards compositions exemplified, for example, by that of boart (see Figure).

The postulated large scale recycling of marine carbonates ($\delta^{13}\text{C} \approx 0\text{‰}$) into the mantle, as suggested by the BABB C + N data, ultimately means that the average composition of 'juvenile' mantle C + N ($\delta^{13}\text{C} \approx -6\text{‰}$ and $\delta^{15}\text{N} \approx -25\text{‰}$) might itself represent a mixture between marine carbonate in pelagic sediments and a more primitive C + N component. This primitive component would be characterized by somewhat lighter $\delta^{13}\text{C}$ and considerably lighter $\delta^{15}\text{N}$ values.

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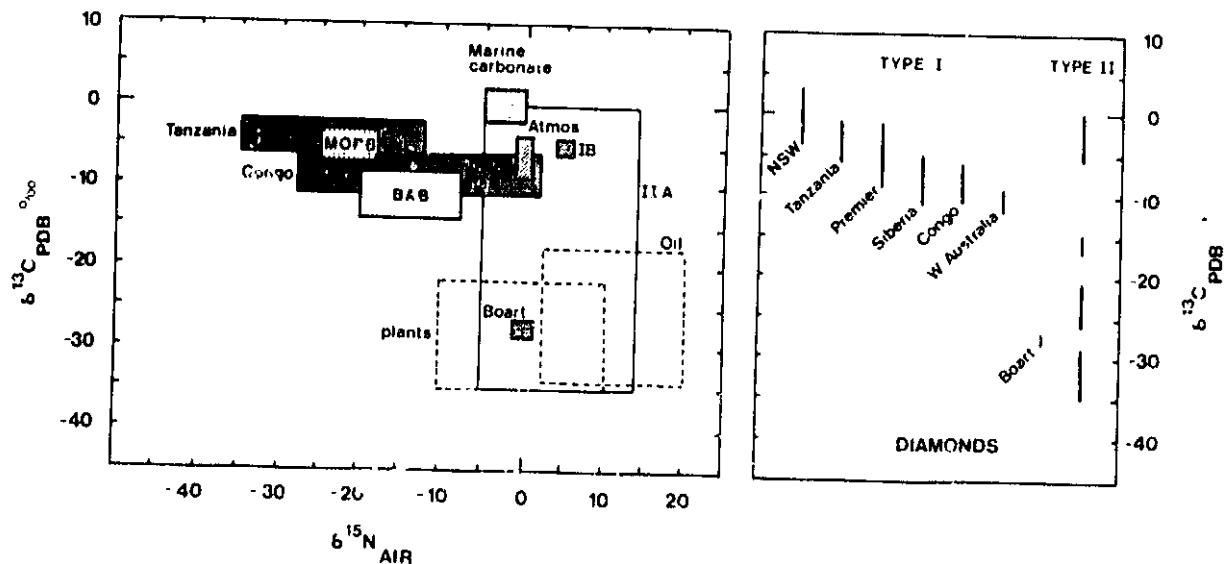


Figure 1 (left). Compilation diagram of carbon and nitrogen isotope data for glasses from mid-ocean ridges (MORB) and back-arc basins (BAB) along with new data for type IA (Tanzania and Congo (Zaire)), IB, IIA and boart varieties of diamond. Composition of marine carbonate, plants, oils and atmosphere are also plotted for comparison. Carbon isotopic variation in diamonds from a variety of localities are compared on the right-hand part of the Figure (NSW - New South Wales, Australia).

CRUSTAL RADIOGENIC HEAT PRODUCTION AND THE SELECTIVE SURVIVAL OF ANCIENT CONTINENTAL CRUST. Paul Morgan, Lunar and Planetary Institute, 3303 NASA Road One, Houston, TX 77058.

The oldest terrestrial rocks have so far revealed no evidence of the pre-approximately 3.9 Ga impact phase of Earth evolution, suggesting that processes other than impact were dominant at the time of stabilization of these units. We are concerned here with the survival of "continental-type" crust in interpreting Earth's oldest rocks, as any early "oceanic-type" crust was subducted within a few hundred Ma of the onset of plate tectonics. Recorded Earth history indicates that while no portions of the continents are completely immune from magmatic and tectonic reactivation, certain areas appear to have more than a random chance of escaping reactivation. With mobile continents, the factors which favor selective survival of certain areas of continental lithosphere must be internal to this lithosphere. Heat flow data indicate that average concentrations of radiogenic heat producing isotopes in surviving Archean crust are relatively low (Figure 1), and this paper investigates a possible relationship between crustal radiogenic heat production and selective survival of ancient continental crust.

A significant portion of terrestrial heat flow from continental regions is generated within the crust by decay of the isotopes K40, Th232, U235 and U238. At present, this part of surface heat flow varies regionally from a few percent to greater than 65% (1). There is no close correlation between heat production and rock composition, except that the heat producing elements tend to be concentrated in lower melting point rocks, and thus, in the upper crust. Surface heat flow, q_s , and heat production, A_o , studies indicate a linear relationship between these parameters in plutonic and metamorphic terranes of the form:

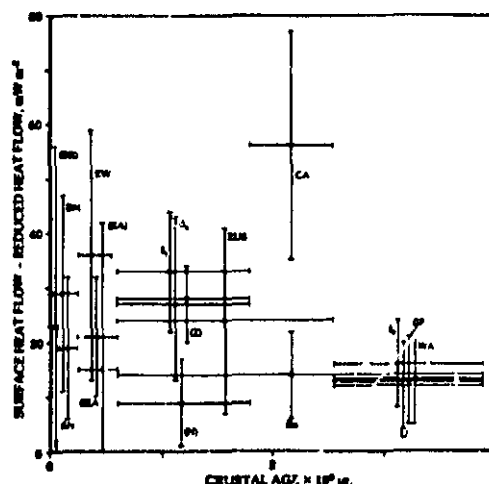


Figure 1. Crustal radiogenic heat production (surface heat flow minus reduced heat flow, an indicator of the average crustal concentrations of the radiogenic isotopes) as a function of crustal age (from ref. 1). Data are grouped according to heat flow province with horizontal bar indicating age assignment of sites in province and vertical bar the standard deviation of the mean surface heat flow in each province. Province codes: BR - U.S. Basin and Range; BM - Bohemian Massif; SN - Sierra Nevada; EW - England and Wales; SEA - Southeastern Appalachians; EA - Eastern Australia; I_1 - Late Proterozoic Indian Shield; Bz - Brazil; N - Niger; Z - Zambia; EUS - Eastern U.S.; CA - Central Australia; Ba - Baltic shield; I_2 - Archean Indian Shield; SP - Superior Province; WA - Western Australia; U - Ukrainian Shield.

Morgan, P.

$$q_g = q_a + A_0 b \quad (1)$$

where q_a and b are constants of the linear relationship for areas termed heat flow provinces (2-6), interpreted to be the heat flow from below the crust, or "reduced heat flow", and a depth scaling parameter for the crustal heat production, respectively. Heat flow from heat generated in the crust is given by the product $A_0 b$, and is typically in the range 5 to 75 mW m^{-2} , although it may be locally higher where associated with granitic plutons.

As crustal heat production is a result of unstable isotope decay, it must have been greater in the past. Assuming the heat producing elements to be immobile after crustal consolidation (time of 1st major magmatic and/or metamorphic activity), their relative concentrations can be calculated as a function of time from the isotope decay constants (e.g. 7). U235 shows the most dramatic decrease in abundance through time, but as the present ratio of U235/U238 is less than 1% (8) the effective combined half-life of the two uranium isotopes is approximately 3.6 Ga. The most significant decrease in isotopic abundance for crustal heat generation over most of earth history is that of K40 with a half-life of approximately 1.55 Ga.

Although heat flow/heat production data sets defining linear relationships of the form of equation 1 allow the total crustal heat production to be estimated, they do not define the distribution(s) of the heat producing isotopes with depth. One of two simple distributions is commonly used; the step distribution in which heat production is constant down to depth b , below which it is zero, and the exponential distribution in which the heat production decreases exponentially with depth, scaled by the parameter b (2,4,9). Regional studies indicate that distributions are more complex and that the different heat producing isotopes may have different distributions and/or depth scaling parameters (10-12). Vertical heat production profiles have been produced for only a very limited number of

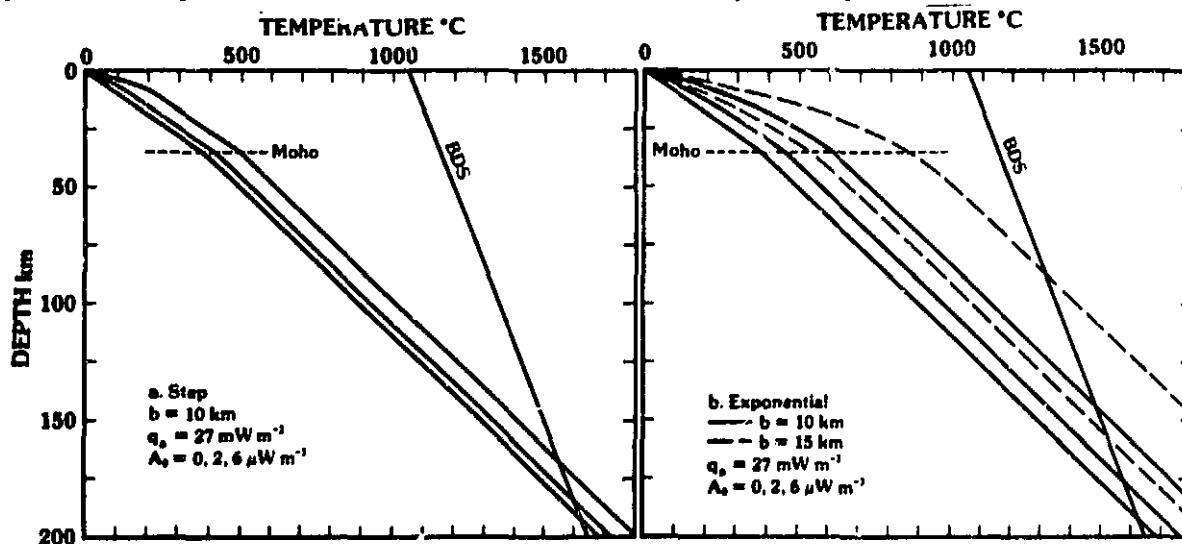


Figure 2. Calculated geotherms for a variety of crustal heat production parameters (from ref. 20). Parameters b , q_a and A_0 are defined by equation 1 above. For each group of curves A_0 increases from left to right as indicated in the figure. BDS is an approximation of the basalt dry solidus. Thermal conductivities of 2.5 and 3.4 $\text{W m}^{-1} \text{K}^{-1}$ were assumed for the crust and mantle, respectively. 27 mW m^{-2} is the present average reduced flow in stable areas. Ranges in other parameters are typical for the continents. Geotherms were calculated assuming the step (left) and exponential (right) distributions of radiogenic isotopes defined above.

Morgan, P.

sites where geological processes have exposed an extensive crustal cross-section. These sites include the Vredefort structure in South Africa (13), the Alps (14), Sudbury (15) and Kapuskasing (16), in Canada, but unfortunately no consistent isotope-depth distribution is evident from these data. The importance for the geotherm of both the total abundance and the depth distribution of heat producing isotopes in the crust is illustrated in Figure 2: Temperatures at depth increase in proportion to the total isotope abundance, but decrease as these isotopes are concentrated upwards. Calculations of the temperature increment at 30 km depth as a function of time from 4.5 Ga to present for a hypothetical crustal section based on data from the Vredefort structure have been given by Morgan and Ashwal (17), and indicate a significantly higher geotherm in the Archean than at present even with no increase in heat flow from the mantle.

The relationship between the geotherm and the stability of continental crust is twofold: Stable crustal thickness cannot exceed the depth at which the geotherm intersects the solidus, and the mechanical strength of the lithosphere is strongly dependent on the geotherm, "hot" lithosphere being weaker than "cold" lithosphere. Crust with high abundances of heat producing isotopes will be remelted at its base with a much smaller increase in mantle heat flow than crust with low abundances, as can be seen from the geotherms in Figure 2. Furthermore, crust with high abundances of radiogenic isotopes will be mechanically weak relative to crust with low abundances, and be more likely to be reconsolidated during tectonic activity. Thus, lithosphere with low heat production crust will be more stable than lithosphere with high heat production crust, and reported differences in relative abundances of K, U and Th between Archean and younger terranes (e.g., 18, 19) may result from selective survival of low heat production Archean crust. Either primary production or secondary differentiation of low heat production crust may be essential for the stability and survival of early continental crust.

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WATER LOSS FROM VENUS: IMPLICATIONS FOR THE EARTH'S EARLY

ATMOSPHERE Steven M. Richardson, Dept. of Earth Sciences, Iowa State Univ., Ames, IA 50011; James B. Pollack and Ray T. Reynolds, Theoretical Space Sciences, NASA Ames Research Center, Moffett Field, CA 94035

The abundance of water on Venus is four to five orders of magnitude lower than on the Earth. This difference may reflect initial differences in the bulk volatile contents of the two planets, or may be the result of processes for massive water loss on Venus. In this study, we perform a series of thermodynamic calculations on the heterogeneous system C-O-H-N-S, varying C/H upward from its terrestrial value of 0.033 in order to evaluate the extent to which either of these two possibilities may account for the low water content on Venus.

We assume that the atmosphere of Venus outgassed rapidly as a result of planetary heating during accretion, and that processes affecting atmospheric chemistry following the period of accretion have been largely those of hydrogen escape and internal chemical re-equilibration. Further outgassing, if any, has had a negligible affect on atmospheric chemistry. In particular, this study investigates the range of atmospheric compositions that can be calculated if we assume that the initial bulk compositions of atmospheres on Venus and the Earth were roughly similar, and that chemical speciation on Venus was controlled by the temperature and oxygen buffering capacity of surface magmas. Our model relies purely on equilibrium thermodynamics and makes no attempt to address those changes in speciation that are kinetically controlled. The atmosphere is assumed to be both vertically and laterally homogeneous for the purpose of these calculations. Petrologic inferences from the Moon (e.g., Warren, this volume) and geophysical models of accretionary heating (Kaula (1) and in this volume) suggest that the surfaces of planetary bodies in the inner solar system were partly or wholly molten during accretion. During this stage, then, we assume that the planetary surface can be modelled as a global magma ocean at a temperature between 1273K and 1573K.

Gases in our model system include but are not restricted to such species as CO_2 , CO , H_2O , H_2 , O_2 , N_2 , H_2S , SO_2 , and CH_4 . At any specified temperature and total pressure, the partial pressures of species in the equilibrium mixture are uniquely defined if the relative molar abundances of carbon, oxygen, hydrogen, nitrogen, and sulfur in the mixture are known. Because an undetermined amount of hydrogen may have been lost from Venus, however, we do not know its abundance in the primitive atmosphere. Furthermore, the total abundance of oxygen in the primitive atmosphere is unknown because the partial pressure of O_2 was controlled by the oxidation state of iron in the molten surface. In order to investigate the range of reasonable initial atmospheric compositions on Venus, therefore, we must place some limits on the proportion of total hydrogen and the buffered fugacity of O_2 . Several estimates of the terrestrial inventory can be found in the geochemical literature. We have chosen to use the one derived by Walker (2). The C/H ratio of 0.033 defines what we consider a limiting case for evolutionary depletion of hydrogen because it corresponds to a system which, in the presence of free iron, would yield a terrestrial mass ratio of water to other gaseous species during accretionary outgassing. Virtually all of the water generated during outgassing under these limiting conditions must later be lost in order to bring the current $\text{CO}_2/\text{H}_2\text{O}$ ratio for Venus up to its observed value of 10^4 to 10^5 . The case for initial hydro-

Richardson, S.M. et al.

gen deficiency on Venus was investigated by using the Walker-derived ratios of terrestrial C/N and C/S, but increasing C/H in successive numerical experiments.

The bulk composition of the outgassed atmosphere is still not defined until we can specify a mole fraction of total oxygen in the atmosphere. It can be seen readily, however, that the relative abundances of all oxygen-bearing species are related to the fugacity of O_2 in the atmosphere through reactions like (1) $2CO_2 = 2CO + O_2$, (2) $2H_2O = 2H_2 + O_2$, (3) $2H_2O + 2SO_2 = 2H_2S + 3O_2$. We have specified the relative total abundances of C, H, N, and S, and although the total relative abundance of oxygen in the atmosphere is not fixed by C/H, C/S, and C/N, we know that it must be equal to the sum of the amounts of oxygen in each of the species in the atmosphere. The fO_2 in the atmosphere, following our basic assumption, is controlled by the oxidation state of the molten planetary surface. A melt produced by accretion of planetesimals, some bearing metallic iron, should be expected to control fO_2 near the buffer assemblage quartz + fayalite + iron (QFI), which controls fO_2 at a value approximately four orders of magnitude below QFM at all temperatures. In order to calculate the abundances of gaseous species at any temperature and pressure of interest, therefore, we specify an fO_2 buffer and calculate an equilibrium fO_2 . Then, using a program derived from a routine by Gordon and McBride (3), we vary the mole fraction of total oxygen in a mixture whose relative abundances C/H, C/N, and C/S have been fixed until the computed species mixture has a partial pressure of O_2 equal to the buffered fO_2 . The mixture calculated by this iterative procedure is our equilibrium atmosphere. For each trial composition, the program uses thermodynamical data from the JANAF (4) tables and minimizes the free energy of a system containing 84 molecular species to arrive at the final mixture.

The proportion of H_2O decreases in model atmospheres with successively higher C/H values, ultimately approaching the "depleted" value currently observed on Venus. The computed ratio CO_2/H_2O increases gradually with increasing C/H, as should be expected, so that bulk compositions used in successive computations reduce the amount of water loss necessary to evolve the present Venusian atmosphere. However, CO_2/H_2O increases roughly as rapidly as C/H. If the present CO_2/H_2O on Venus of 10^4 to 10^5 is not due to water loss, but is due instead to an initial hydrogen deficiency relative to the Earth, then that deficiency must have been on the order of 10^5 to 10^6 . The dynamics of mixing in the inner solar system during accretion make this difference between Venus and the Earth unlikely (5).

Increasing C/H also results in a rapid increase in CO/H_2O , however, and provides an efficient mechanism for water loss by the reaction $CO + H_2O = CO_2 + H_2$. This reaction, together with water loss mechanisms involving crustal iron, could have removed a very large volume of water from the atmosphere of Venus, even at rather low values of C/H. In fact, we calculate that a C/H difference of a factor of 5 or less between Venus and Earth could have been sufficient, under the fO_2 conditions that may have prevailed during early outgassing, to produce the observed difference in water content between the two planets by the CO reaction alone. We do not claim that the CO reaction was solely responsible for the present low water abundance on Venus, but choose to emphasize that the difference between the Earth's outgassed C/H value and one for which we calculate roughly equal amounts of CO and water is relatively small. By an accident of bulk chemistry, Earth may have narrowly avoided water loss similar to that on Venus.

Richardson, S.M. et al.

Selected Molecular Species in

Gas Mixtures Calculated at 1373K, 100 atmospheres, and $fO_2 = QF1$

	<u>C/H = 0.033</u>	<u>C/H = 0.17</u>	<u>C/H = 0.325</u>	<u>Observed^a</u>
CO	4.25 atm	19.82 atm	32.14 atm	1.9×10^{-3} atm
CO ₂	0.80	3.73	6.05	91.675
H ₂	67.45	51.76	40.64	10^{-3}
H ₂ O	25.94	19.95	15.67	$< 10^{-2}$
N ₂	0.18	0.80	1.02	3.8
CH ₄	1.19	3.24	5.25	-----
H ₂ S	0.15	0.62	0.96	2.85×10^{-4}
CO/CO ₂	5.31	5.30	5.30	2.1×10^{-5}
CO/H ₂ O	0.16	1.0	2.05	0.2

^a calculated from Von Zahn et al. (6)

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CHEMICAL EVOLUTION OF THE EARTH--EQUILIBRIUM OR DISEQUILIBRIUM PROCESS?
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In discussing a homogeneous accretion scenario for the origin of the Earth, most authors agree that gravitational differentiation played a major role, but disagree about whether the core formed in equilibrium with the mantle. The disequilibrium hypothesis arose from the recognition that the upper mantle is too oxidized and contains excessive abundances of siderophile elements to have segregated metallic iron [1]. Abundances of other elements have also added to the puzzle. To explain the apparent chemical incompatibility of the core and the mantle, various core-forming mechanisms have been proposed. These mechanisms include rapid disequilibrium sinking of molten iron [1,2], oxidized core or protocore material [3, 4], and meteoritic contamination of the upper mantle after the separation of the core [5]. I believe that we should examine the meaning of "chemical equilibrium" in a planetary interior before we discuss the pros and cons of these alternative mechanisms.

In discussing "equilibrium", most authors have addressed only equilibrium in a local environment where segregation or passage of proto-core material presumably took place, and their discussions were based on a one-way traffic system. They focused attention on the sinking of iron alone, ignoring the fate of the matter such as volatile elements that presumably resided in the deep interior. Investigators of the origin of the atmosphere and hydrosphere, on the other hand, have shown interest only in the ascent of volatile elements, primarily after the formation of the core. I see no valid reason to exclude the concurrency of the two processes. The deep interior could have been heated to temperatures high enough for many volatile elements to start ascending by the time the hot protocore material, initially dispersed throughout the Earth, began accumulating in the core region. Some evidence even suggests that the Earth was entirely molten shortly after accretion [6]. If the Earth was once molten, elements (including iron) would have been reasonably free to move in all directions. An important question is how the direction of transport of each element is determined in a two-way traffic system.

Except for the generation of heat by radioactive decay, a spontaneous process takes place in the direction of a lower energy content within a closed natural system. A system is said to be in chemical equilibrium when it attains the lowest chemical energy state under a given set of physical parameters of state (P, T, gravitational acceleration, etc.). In this state, net transfer of matter among phases ceases. In petrological discussions, a system is usually chosen to have such small dimensions that the physical parameters can be considered uniform. The current equilibrium-disequilibrium controversy is based on the definition of equilibrium in a uniform system. When we try to discuss the spontaneous redistribution of elements in a planetary interior (i.e., chemical evolution), such discussions are inappropriate. Instead, we have to treat the whole global system as a continuum system, in which gradients exist in the parameters of state, and search for the condition under which no net transfer of matter would take place in the radial direction (static equilibrium state).

An early attempt to assess stable distribution of elements in the Earth was made by Brewer [7], who pointed out that the effect of gravity on chemical potential is such that iron should be concentrated in the core region and oxygen in the outer regions, and that uranium is concentrated

in the crust because of its strong chemical affinity to oxygen in spite of its high density. Because of theoretical difficulty for classical thermochemistry in dealing with nonisothermal processes, he did not go beyond a conceptual presentation. By adopting concepts used in steady-state thermodynamics, I have devised a method for evaluating how elements should distribute stably in the Earth's interior for the present gradients of temperature, pressure, and gravitational acceleration [8]. The method consists of (1) calculating fugacities of elements (in the ideal gas state) for assumed phase assemblages at various depths using the free energies of formation and molar-volume values of phases at T and P along an assumed geotherm, and (2) correcting the fugacities for the barometric effect. The corrected fugacity is called equigravitational fugacity (the symbol: f^*) [8]. The first step takes into account the chemical energy involved in forming the phases and the effect of pressure in modifying the relative stabilities of phases. The second step makes corrections for the contribution of gravitational potential to chemical potential. The spontaneous direction of chemical evolution is toward the uniformity of the equigravitational fugacities for each element. The method was derived on the basis of the following considerations: (1) a static condition is obtained when fugacities are uniform across a temperature gradient in a system in which heat flows steadily; (2) lithostatic pressure increases the fugacities of elements in a condensed phase as a function of molar volume; and (3) the barometric relation caused by the presence of a gravitational field requires that the fugacity of an element increase with depth to maintain a static state. The barometric relation for a global system in which both temperature and gravitational potential vary with depth is given as

$$\ln (f_2/f_1) = (M/R) \int_{z_1}^{z_2} (g/T) dz \quad (1)$$

where f_1 and f_2 are fugacities at depths z_1 and z_2 , M is the molecular mass of the gas species, R is the gas constant, g is the gravitational acceleration, and T is the absolute temperature. In a single-element global body that maintains an adiabatic temperature gradient, the above factors interact in such a way that the corrected fugacity of the element remains uniform. In a multi-element body, a compositional differentiation in the radial direction can be expected because the barometric relation depends on the molecular weight of the species involved (barometric differentiation).

Thermochemical modeling based on the above method was performed for three temperature regimens (i.e., Fe-FeS eutectic, present Earth, and peridotite melting) and for both homodensity and density-stratified Earth models. The Fe-FeS eutectic geotherm represents the minimum temperature regimen in which iron can move to the core as liquid [9, 10]. The present geotherm is approximated by the equation: $T = -1462 + 1432 \log (Z + 17)$, where T is the absolute temperature and Z is the depth in kilometers. This equation gives 300 K at the surface, 1,500 K at 100 km, and 3,500 K at the core-mantle boundary. The peridotite-melting geotherm represents the thermal condition in which elements are relatively free to move in all directions.

The results obtained to date show the following.

(1) For all temperature models, the equigravitational fugacities of iron and oxygen are much less uniform for a homodensity Earth than for a density-stratified Earth, indicating that a homogeneously accreted Earth was highly unstable chemically regardless of whether the Earth accreted hot or cold.

Sato, M.

(2) The direction of spontaneous chemical evolution for a homogeneously accreted Earth is for metallic iron to be concentrated in the core region and oxygen in the outer regions, as conceptually suggested by Brewer [7].

(3) For both homodensity and density-stratified Earth models, the $f_{\text{Fe}}^{\text{Fe}}$ becomes more uniform as the temperature increases. The $f_{\text{Fe}}^{\text{Fe}}$ distribution in the density-stratified Earth approaches uniformity fairly closely in the peridotite-melting temperature model, suggesting a possibility that the present distribution of iron in the core and the mantle was established shortly before the solidification of the mantle.

(4) In the present Earth, the $f_{\text{Fe}}^{\text{Fe}}$ distribution is not uniform, even if the activity of FeO in the mantle is as low as 0.1. This $f_{\text{Fe}}^{\text{Fe}}$ value is greatest at a depth of about 450 km. This finding suggests that the present Earth is not in the static equilibrium state, and that iron should move down to the core region below this depth and, at the same time, should move up to the surface as oxides above this depth. The latter requirement appears to be met by magmatism.

(5) If the oxygen fugacity of the mantle is assumed to be buffered by the core-mantle boundary, which is the most potent buffer assemblage in the Earth, the mantle should be increasingly oxidized as the radial distance from the core-mantle boundary increases. If the activity of FeO at the core-mantle boundary is assumed to be 0.1, the upper mantle at a depth of 100 km should have an oxygen fugacity within the stability field of magnetite at static equilibrium. This finding contradicts the contention of Ringwood [1] that the upper mantle is too oxidized to be in equilibrium with a metallic core. Perhaps the upper mantle is not oxidized enough to be in static equilibrium with the metallic core. Magnetite is not common in mantle samples.

(6) The present degassing of primordial ^3He may be due to the increased barometric differentiation caused by the cooling of the Earth.

Thermochemical modeling seems to give us useful insights into the nature of chemical evolution of the Earth without invoking overly speculative assumptions. Further work on siderophile elements, rare gases, and possible light elements in the outer core is warranted. Because the method is so general, it is also a useful tool for comparative planetology.

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LARGE ARCUATE SCARS: A GEOLOGICAL LEGACY OF THE EARTH'S ACCRETIONARY PAST, John M. Saul, ORYX, 3 rue Bourdaloue, 75009 Paris, France.

Immediately following accretion, the surface of the Earth was densely patterned with circular scars which were the surface expressions of 3-D craterform structures. In the course of geological time these structures would have become less and less visible due to the workings of the Earth's atmosphere, surface waters, and plate tectonics regime but there is no compelling reason to assume that they have been entirely eradicated. Furthermore, a very imperfect analogy with the other inner planets suggests that geological processes may not in fact be capable of totally erasing such deep features, though an equally imperfect analogy with Io can be used to argue otherwise. Two opposing points of view are thus possible, that the Earth's present surface and crust contain traces of its earlier pock-marked appearance or, alternatively, that all indications of such markings are long gone.

The Earth's crust is brittle down to a certain depth; below this it is ductile. Fractures situated entirely in the brittle domain are susceptible to closure and annealing but the lower extremity of any fracture extending all the way down to the brittle/ductile transition is subjected to horizontal and vertical movements from the adjacent ductile materials. On a physically and thermally active planet such fractures would remain open over great periods of time, perhaps indefinitely, and through large vertical extents. Younger rock units, whether overlain as surface sediments or igneous extrusives, or inserted as intrusives, would be subject to fracturing from below at their contacts with the active fracture zones here envisaged. The time scales of geological processes on and near the Earth's surface will in general be entirely independent of the time scale involved in movements within the ductile region far below. In consequence, some deep arcuate fractures should be covered over and temporarily invisible at any particular geological moment, whereas in other areas they may be more or less detectable. Thus the search for traces of the Earth's accretionary past should not be a search for deeply-eroded or annealed craters, or shocked rocks, but rather for large arcuate structures or fracture zones.

The large sizes which we can safely assign to some of the early impact scars, combined with their hypothesized survival over the subsequent $\sim 4 \times 10^9$ years, suggest that many features attributable to the Earth's accretionary period should also be of large scale. Meteorite fragments, kink banding, shatterconing and similar small-scale features could not have been preserved by the process of sequential inheritance proposed here but features of large horizontal extent such as mountain ranges, valleys, faults, rifts and shields might. Similarly, volcanism and certain categories of hydrothermal mineral deposits may indicate crustal weakness and openings over sufficient depth to have been preserved. In general we should look for zones of circular curvature (with fracture-enhanced permeability) whose visibility has been further developed by the action of rising or descending fluids. Over a thousand circular scars have been identified during the last decade.

A selection of illustrative or spectacular examples includes:

Arizona: An area in Arizona bounded by latitude 33° - 35° N and longitude 110° - 112° W contains a minimum of 19 circular scars, ranging in diameter from 18-110 km.¹ The largest circle is particularly easy to see and its perimeter from the 12 o'clock position to the 2 o'clock position is defined by the Mogollon Rim. In this area there are 24 known mineral deposits whose mode of origin appears to involve the movement of fluids and most of these lie in the 9% of the total area classified as "on" a rim. These circles were originally

SAUL, J.M.

detected on raised-relief maps sprayed white, illuminated at a low angle, and photographed from above.

Lake Victoria: To the southeast of Lake Victoria the gentle curvature of Lake Eyasi and its anomalous northeast-southwest orientation permit the eye to pick up a smooth continuous arc, 90° in length, composed of lakes, volcanoes, craters, hot springs and faults in the Eastern Rift. This arc brackets Lake Victoria from approximately the 2 o'clock to the 5 o'clock position. (Radiometric ages obtained from northeast-southwest trending dolerites at Lake Eyasi give three clusters around 2500, 1900 and 900 million years,² strongly suggesting the occasional rejuvenation of ancient fractures rather than some fleeting or Recent geological happenstance.) On the other side of Lake Victoria, directly opposite Lake Eyasi (i.e. at the 11 o'clock position), the poorly-understood Bunyoro Series of Precambrian metasediments has an orientation, width and puzzlingly curved axis³ which, in conjunction with lithological considerations, lend themselves to a provisional interpretation as a sedimentary unit derived from an adjacent arcuate highland. To the south and west some of the volcanic and lacustrine features of the Western Rift appear to follow the curvature of the Bunyoro Series but below the 9 o'clock position the Western Rift diverges westwards away from Lake Victoria. An overall impression is that of a very large linear or near-linear fracture propagating from the north-northeast, abutting perpendicularly against a resistant circular plug (diameter ~650 km) whose center, a topographic low, is occupied by Lake Victoria. Not being able to penetrate the plug, the fracture bifurcated and, taking two symmetrical paths of low resistance, wrapped itself around it. This inability of a rift-like fracture to cut through a circular plug-like structure has implications in the interpretation of other regions of long-term geological stability as well.

The Canadian Shield, for example, can be interpreted as corresponding to a circular scar (3700 km in diameter), western portions of which are incidentally covered by sedimentary formations, that has maintained its integrity throughout geological history. The Black Hills and the Adirondacks, two distinctive areas of anomalous uplift, sit on its rim zone.

The arc-shaped Aleutian trend intersects the Kurile-Kamchatka arc in the volcanic region of Mt. Shiveluch. Topographically or geophysically mapped, this intersection takes on a stubby, tapering cusp shape, or just what might be expected by the three-dimensional intersection of two deep craterform scars of circular surface expression. Circles, however, must intersect in two points, not one, and the other point of intersection of these arcuate trends extended is on (or very near) Bennett Island in the East Siberian Sea, site of a major volcanic event since February 1983.

Geometrically, the original impacts were craters with steep walls whose sides flattened out with depth. The fractures in lower portions of the brittle crust will have been at comparatively shallow angles and these angles would have remained the same throughout geological time. This, however, would not have been the case for the steeper, higher portions of the crater walls. When eroded and subsequently covered by younger rocks the upper portions of crater walls would have been "inaccurately" regenerated from below at shallower angles, determined, not only by the original crater morphology, but also by the maximum depth of erosion. In the course of time then some originally steep arcuate fracture zones would have become substantially less steep, eventually, as in the case of the Aleutian Trench, attaining the more moderate angles of descent preferred by subducting plates. This is consistent with the observation that oceanic trenches occur on the convex sides of island arcs and with the opinion held by many geologists that

SAUL, J.M.

the subduction of tectonic plates did not commence until relatively late in the Precambrian.

These matters are all clearly more complicated than this outline indicates. The curvature of the Earth plays a non-negligible role. Some circles exhibit radial as well as circumferential fractures. Rocks normally tend to break in straight lines, not curves. Central peaks and higher order (outer) scars can be detected in some instances but not others. Furthermore, the three-dimensional geometry of intersecting craterform scars with modified angles of descent can be exceedingly difficult to reconstruct. But in general this type of thinking seems to provide useful ideas concerning the canvas on which later geology has been painted.

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LUNAR AND MARTIAN IMPACT BASINS: EXPOSED RECORDS OF TERRESTRIAL BOMBARDMENT? Peter H. Schultz, Lunar and Planetary Institute, 3303 NASA Rd. 1, Houston, Texas 77058

The unrecycled surfaces of the Moon, Mercury, and Mars preserve the very early history of impact bombardment and its effect on crustal evolution. Previous studies indicated that the post-accretion impact flux by large bodies on Mars may have been deficient, but systematic studies (1) recently have revealed that this deficiency is largely the result of active erosional and depositional processes during the first 0.8 by. Ancient martian impact basins larger than 300 km in diameter are revealed by subtle but unequivocal topographic and structural control of drainage patterns. In addition, the largest basins (> 2000 km in diameter) have left a deep-seated imprint of concentric and radial structural patterns that control the occurrence of most martian volcanic and tectonic provinces (2).

The areal density of lunar impact basins larger than 300 km is about 10^{-6} /km² (3). Newly recognized large basins on Mercury (4) now bring the tally for that planet close to 1.6×10^{-6} /km². Although numerous new basins have been found on Mars, the areal density remains lower than that of the Moon and Mercury at 0.5×10^{-6} /km². This apparent deficiency is almost certainly the result of relatively conservative criteria used for identifying martian basins (e.g., multiple rings) that exclude the numerous arcuate but incomplete basin-like rims. If 10^{-6} /km² is considered to be a reasonable approximation for the Moon and Mars, then the Earth should have recorded more than 500 impacts that resulted in basins larger than 300 km in diameter over its post-accretion geologic history. If calibrated with the Moon, then most of these impacts occurred prior to 3.8 by.

Each planet appears to preserve one mammoth impact structure: Imbrium on the Moon (1200 km in diameter); Caloris on Mercury (1300 km); and Hellas on Mars (2000 km). Although larger basins have been reported, the ring diameters probably correspond to structural modifications that can not be directly compared to the better preserved examples listed above. Nevertheless, each of these basins cover between 53 and 69% of the planetary radius, and the calculated impact energy approaches 0.1% of the gravitational potential energy of the planet. The response of the crust and mantle to the large transient cavity results in the uplift of the crust-mantle boundary revealed by positive gravity anomalies (5).

Impact basins on the Moon and Mars largely control the distribution of post-accretion volcanism and subsequent structural history. Radial grabens extend more than 2000 km from the Imbrium basin, thereby affecting more than 70% of the nearside hemisphere (6). Similar radial fractures have recently been reported for the major martian basins (2). Massive eruptions of basalt were localized along concentric zones of weakness that also extend well beyond the nominal raised basin rim. On Mars, concentric canyons and grabens encircle the Hellas basin 2400 km from basin center; that is, the concentric structural zone of influence covers nearly 80% of one hemisphere. The continued re-use of these structural weaknesses as volcanic conduits throughout the geologic history of both Mars and the Moon underscores their importance for the evolution of the early crust on these planets and perhaps provides a clue for the role of major impacts for evolution of the early Earth.

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THERMAL, DYNAMIC AND COMPOSITIONAL ASPECTS OF THE CORE-FORMING EARTH. D.J. Stevenson, Division of Geological and Planetary Sciences, Caltech, Pasadena, CA 91125.

Core formation is the most important and singular differentiation event in the history of a terrestrial planet. It almost certainly involved the downward migration of a partially or wholly molten iron alloy through a silicate and oxide mantle, and was contemporaneous with accretion [1]. (There are many arguments against direct accretion of the core from iron rich planetesimals, one of which is the inability of such a directly formed core to be consistent with the properties of the present core, i.e., mostly molten, leavened with one or more light elements.) I deal here with several important, unresolved issues which have implications for mantle and core geochemistry, the thermal history of the earth, and the origin of geomagnetism: Was the early earth molten? Does core formation involve low or high pressure geochemistry, or both? Can the early earth mantle homogenize? Is equilibration established between core-forming material and the mantle through which it migrates? How much iron is stranded and unable to reach the core (i.e. are mantle siderophiles such as Mo explained by late infall or by stranded iron?)

Was the Early Earth Molten? A magma ocean figures prominently in most speculations for the early moon (see [2], p.242 onwards) despite indications that it is neither physically plausible [3] nor geochemically necessary [4]. The putative terrestrial magma ocean [5] is energetically more plausible since accretion supplies much more than enough energy to completely melt the earth, yet still dynamically questionable because the earth may be able to "buffer" i.e. get rid of accretional energy as quickly as it is supplied. Buffering by vigorous convection in a partially molten medium is conceivable [6] but cannot be quantified with confidence because the appropriate parameterization of this convection does not exist yet (numerical simulations are needed). Consider, instead, a model in which extensive partial melting causes frequent eruptions and surficial lava lakes which form chill crusts. If the earth formed in $\sim 3 \times 10^7$ yr (the usual estimate, ref. 7) from planetesimals ~ 10 km in radius then there are ~ 10 impacts/yr and the time between disruptions of the chill crust by either direct impact or ejecta would be $1-10^2$ yr. In this time, the crust has grown to a thickness of 2-20 meters and the average heat outflow has been $3 \times 10^5 - 3 \times 10^6$ erg/cm²-s (equal to σT^4 for $T \sim 270-480$ K, so most of the loss is determined by conduction through the crust rather than direct radiation). By comparison, the time and space-averaged energy input of accretion is 5 to 10×10^4 erg/cm²-s (assuming 10-20% energy retention, the rest being promptly radiated during the impact events). Since the thermal energy loss rate can exceed the input, I conclude that there is no compelling reason to expect a magma ocean. Indeed, even $\sim 10\%$ partial melt is highly mobile and may suffice, although a higher degree of melting (e.g. 25% as in Anderson's model [8]) certainly cannot be excluded and might be expected by pressure-release melting in upwelling plumes.

What is the Nature of the Core-forming Fluid? The physical processes have been discussed elsewhere [1] and I focus here on the geochemical question of whether core-forming alloys form at high or low pressures. Because of the nature of accretional energy deposition, the most likely place to macrosegregate core-forming melt is always near surface (e.g. at the base of the outer thermal boundary layer). The only potential counterexample is

Stevenson, D.J.

the model proposed by McCammon et al. [9] in which the Fe-FeO phase diagram exhibits a markedly non-monotonic dependence of eutectic temperature on pressure so that an Fe-O melt can form in the lower mantle only. However, iron must be convected into this high pressure region and be present in disseminated form in order that their model work. A more probable scenario is low pressure macrosegregation and rapid diapir-like descent of an alloy containing Fe, Ni, S, possibly H [10], perhaps C, N, Si, but only a small amount of oxygen. Another interesting consequence of this is the oxygen fugacity conditions below the outer boundary layer are likely to be similar to those in the present earth mantle.

Equilibration and Homogenization. The diapiric downwellings of iron have characteristic dimensions of kilometers or larger [1] and are unable to equilibrate efficiently with the partially molten material that they traverse. Metal-silicate equilibration only occurs in those regions where the iron remains in disseminated form (droplets, veins, grains). This is most probably in the near surface region where the iron is being continuously resupplied in disseminated form by impacts and undergoes downward percolation and macrosegregation. However, the fluid dynamics is complicated and one cannot exclude the possibility of significant entrainment of disseminated metallic iron to deep mantle regions (just as particulate iron can be entrained upward by mafic magmas, as in the Disko Island basalts, West Greenland [11]). One also cannot exclude Rayleigh-Taylor instabilities ("subduction") of iron-contaminated silicate blobs which migrate to the core-mantle boundary.

Sub-solidus convection is incapable of homogenizing material down to the centimeter lengthscale, but the combination of this large scale circulation together with the percolative action of partial melting may achieve homogenization provided stable stratification does not develop. It seems likely, therefore, that despite the disequilibrium between core and mantle, the mantle may either wholly or within discrete layers establish a remarkable degree of equilibrium and homogenization. The existence and number of mantle layers is determined by both the degree of heterogeneity of accretion (specifically the variation in major chemistry between early infalls and late infalls) and the degree of macrosegregation from partial melting. It is possible that homogeneity and whole mantle convection occurs despite a gradual evolution in the density of incoming material, even if this evolution is in the sense of creating a stable density gradient. This occurs if $\Delta\rho_x \leq \Delta\rho_T$, where $\Delta\rho_x$ is the total density reduction in the outer region due to admixing of lighter material in \sim one convective overturn time and $\Delta\rho_T$ is the density difference driving the convection. Crude models suggest that $\Delta\rho_x \sim 10^{-3} \text{ g.cm}^{-3}$, so that the inequality is satisfied. However, this simple argument can be negated by single, very large impacts late in the accretion or by massive macrosegregations.

How Much Iron is Stranded? Since core formation is surely not 100% efficient, some iron (and the siderophiles that partitioned into it) should be stranded in the mantle. This may be relevant for understanding mantle siderophile and volatile abundances. Consider percolative (D'Arcy law) flow of iron melt along microscopic channels between silicate grains. For 1 mm grain size, the melt velocity is $\sim 50f \text{ cm/s}$, where f is the total volume occupied by the melt. In a quiescent mantle that gradually allows melt to drain from it, one finds that

Stevenson, D.J.

$$\frac{df}{dt} \approx \frac{-50r^2}{H}$$

in a simple model in which f is constant throughout a mantle of depth H . Solving, one finds that after an initial transient period, the value of f at $t = 0$ is unimportant and $f(t) \approx H/50t$. Consequently, f drops to 0.01 in only 10^2 yr or less! There are at least two fallacies in extending this argument to very low melt fractions. First, it does not allow for channel closure due to surface tension effects, expected to occur at some low melt fraction (10^{-2} - 10^{-3}). Second, it does not allow for the "suspension" or freezing of iron as the mantle overturns. For these reasons, no accurate prediction is possible, but significant stranding is expected.

What are the Outstanding Issues? The thermal state of the early earth remains poorly constrained despite the arguments presented here. Clearly, one would like to know the composition of the core, the nature of the inner core, the composition and degree of homogeneity of the deep mantle, whether the moon has a core, and the extent to which mantle abundances are consistent with low pressure metal-silicate fractionation. The composition of the core, in particular, may continue to resist a definitive determination.

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EARTH'S PRIMORDIAL DIFFERENTIATION, AND ITS AFTER-EFFECTS

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The case for massive primordial melting on the Earth is circumstantial (i.e., based mainly on lunar evidence), but nonetheless compelling. The term "magma ocean" is probably a hyperbole. Among other things, "ocean" implies the system is virtually 100% liquid, with a mainly gas/liquid upper surface, and a water-like viscosity of the order 10^{-2} poise. I prefer "magmasphere."

1. EVIDENCE FOR A LUNAR MAGMASPHERE

The case for a primordial lunar magmasphere is strong, although the degree to which this system resembled an "ocean," and its relationship to the many "Mg-rich" types of ancient lunar cumulates, are certainly debatable [1].

(1) Based on three fairly typical highlands landing sites (Surveyor 7, Luna 20 and Apollo 16), the region that spawned the ALHA81005 regolith meteorite, and remote sensing data [2] for Fe, Mg, Al/Si, and Mg/Si in the regolith extending over two broad, low-latitude bands (which, for Fe and Mg, are circumglobal), the upper lunar crust averages about 75% plagioclase. The anorthosite layer is truly global: the only large areas where the regolith plagioclase content is $<75\%$ are covered by mare basalts, which appear to be veneers generally <2 km thick [3]. The thickness of the plagioclase-enriched layer can be constrained by gravity data (which indicate "mascons" beneath circular basins, and conversely, low densities for the crust elsewhere) to be at least 30 km [4]. The upper crust is enriched in Al over the bulk-Moon composition (constrained by various means, e.g., heat flow [5]) by at least a factor of 5. The (rare) samples of anorthositic lithologies that survived the meteoritic bombardment in "pristine" form generally have cumulate textures. The simplest, most plausible mechanism for producing a near-surficial, global high concentration of plagioclase is flotation as a cumulus phase over a denser magmasphere.

(2) The average age of eight pristine cumulates dated by Rb-Sr and/or Sm-Nd internal isochrons is about 4.32 Ga [1, 6]; none is <4.08 Ga, and three are >4.45 Ga. Model ages for younger extrusive lithologies are generally similar (4.3-4.4). Clearly, a cataclysmic differentiation, which produced at least half of the present crust, occurred during the first 200 Ga of lunar history; and the Moon became relatively quiescent after the first 400 Ga. (3) There is also the well-known argument about Eu anomalies, in which respect mare basalts and the average highlands appear complementary [7]. (4) The composition of KREEP is remarkably uniform over the sampled portion of the Moon [8], which suggests that it is all of common origin, as a residual liquid from a single global magma. (5) Patterns of petrochemistry correlated with longitude [2, 9] also suggest a global rather than localized differentiation process.

The main reason for doubting that the Moon had a magmasphere is that no mechanism is known for heating the Moon rapidly enough to make it more than a few percent molten. This problem is effectively avoided by Shirley's [10] "partially molten magma ocean" model, in which most of the magmasphere is a "magmifer" that is fluid and convective, but only partially molten; only a relatively thin layer at the top is fully molten. Shirley argues that convection inevitably works to cool the system enough so that the bottom can never be more than a few degrees above the solidus (i.e., a few percent molten). Convection also works to maintain an adiabatic P-T relationship (about 4 K/kb, for a partially molten system). In contrast, silicate melting curves have $dT/dP = 10$ K/kb. Thus, the percentage of melt increases toward the top of the "magmifer," which occurs at roughly -25 kb from the base of the system, where the fraction of melting (a function of $T - \text{melting } T$) is such that melt

Warren, P. H.

buoyantly separates from crystals, and rises into the fully molten layer at the top of the system. This model accounts for massive differentiation of the outer few hundred km of the Moon, without requiring an implausibly high input of primordial heat.

2. RELATIVE A PRIORI PROBABILITY OF FORMATION OF A MAGMASPHERE ON THE EARTH

A priori, a magmasphere is more likely on the Earth than on the Moon, for numerous reasons: (1) Mass/surface area is $6.1 \times$ greater for the Earth than for the Moon. Therefore, other factors being equal, the Moon cools faster, and is less likely to build up sufficient heat to produce a magmasphere. (2) The various heat source(s) proposed to account for the Moon's magmasphere probably were comparably efficient for the Earth [11]. In addition, conversion of gravitational energy into heat during infall of the Earth's large core provided enough heat to raise the planet's mean temperature by as much as 2300 K (the corresponding value for the Moon is 10 K) [12]. (3) The Earth's greater bulk also gives it an escape velocity $4.7 \times$ greater than that of the Moon, which implies that the Earth's capacity for retaining atmospheric gases is vastly greater. A thick, dense proto-atmosphere probably insulated the primordial Earth's surface [13]; the Moon probably also had a primordial atmosphere [13], but it must have been relatively tenuous, and less efficient as insulation. (4) It has been suggested [14-15] that because densities of silicate melts are possibly higher than ultramafic rock at pressures greater than roughly 100 kb (i.e., depths below roughly 300 km in the Earth, but nowhere in the Moon), heat build-up in the lower mantle might be facilitated.

3. EVOLUTION OF THE EARTH'S MAGMASPHERE

A magmasphere in an Earth-sized planet is different from a magmasphere in a Moon-sized body. The median pressure in the non-core portion of the Earth is 410 kb; the corresponding value for the Moon is only 17 kb. As discussed above, input of primordial heat/mass was probably comparable for the Earth and the Moon (if anything, higher for the Earth). If equal percentages of the non-core portions of the Earth and the Moon were partially melted and incorporated into their respective magmaspheres, the pressure range in the terrestrial magmasphere was roughly $20 \times$ that in its lunar counterpart. For example, a terrestrial magmasphere with a 600 kb pressure range (1500 km depth range) is commensurate with a lunar magmasphere with a 25 kb pressure range (500 km depth range). As discussed above, assuming (in accordance with Shirley's [10] lunar model) that the lower, "magmifer" portion of the system has an adiabatic (4 K/kb) P-T profile, it gives way to a fully molten layer roughly -25 kb (70 terrestrial km) above the bottom of the system. Thus, the terrestrial magmasphere probably had a relatively thick fully molten layer, and a relatively thin magmifer layer, in comparison to its lunar counterpart.

Heat transfer (by convection) is in general far more rapid in a fully molten layer than a relatively viscous "magmifer." Therefore, the heat input required to increase the total system (magmifer layer plus upper, fully molten layer) thickness beyond the maximum magmifer thickness (roughly 25 kb) is far greater than that required to thicken it up to the maximum magmifer thickness. The system tends to self-regulate, such that the thickness of the total system is unlikely, barring overwhelming heat input, to ever become much greater than the maximum magmifer thickness. As a result, the percentage of the non-core portion of the Earth that was partially melted and incorporated into its magmasphere was probably considerably smaller than the corresponding percentage for the Moon, other factors (discussed above) in favor of greater primordial heat input and retention for the Earth notwithstanding.

It might be assumed that if the Moon's magmasphere generated a thick anorthositic crust, a terrestrial magmasphere would have acted likewise. The amount of anorthosite produced by the Earth's magmasphere is difficult to

Warren, P. H.

constrain, but it clearly was far less than commensurate with the amount of primordial lunar anorthosite. This disparity is probably mainly due to the greater pressures in the Earth's interior, compared to the Moon's interior. Pressures in the Earth's magmasphere led to garnet crystallization at depth, leaving little Al to form feldspar near the surface [16].

4. AFTER-EFFECTS

Anderson [17] suggests that a terrestrial magma ocean produced a layer of eclogite in the mantle between roughly 300 and 670 km depth, and that this layer is the source region for mid-ocean-ridge basalts. There is considerable evidence that upper mantle convection extends down to at least 670 km (e.g., [18], in which case it is unlikely that a compositionally distinctive layer could persist for more than a few convective cycles. The mantle above 670 km is probably modestly heterogeneous, but the heterogeneity is probably as much horizontal as vertical.

Hofmeister [19] has shown that Archean sialic crust was probably not produced directly from a relatively shallow (120 km) magma ocean. However, the discrepancy between the composition of the primordial crust predicted in her model and the actual composition of the Archean crust would have been smaller if she had modeled a magmasphere thicker than 120 km, due to pressure effects on pyroxene/olivine stability (important for Ca and Si; cf. [16]), as well as simple mass balance considerations (for potassium). It is possible that a major portion of the Archean sialic crust was produced by intense metamorphism of a proto-crust formed atop the magmasphere.

The most important direct evidence in favor of a terrestrial magmasphere is probably the isotopic evidence [20] for massive degassing of ^{129}I (with a half-life of 17 Ma) from the upper mantle. As discussed in detail elsewhere [21], large regions of the Archean upper mantle probably remained H_2O -poor for several Ga after the primordial degassing of the magmasphere. The resultant higher (by roughly 250 K) upper mantle melting temperatures caused these regions to have higher viscosity, and thicker (by nearly a factor of two) lithosphere than would otherwise have been the case. Production of ultramafic komatiites was facilitated, for several reasons: (1) temperatures were higher (by roughly 250 K) in the asthenosphere, and (2) pyroxene stability was enhanced in the anhydrous komatiite source regions, resulting in higher MgO in partial melts per given f , and lower dT/df during melting.

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CHEMICAL CONSTRAINTS ON THE EVOLUTION OF
ARCHAEAN CONTINENTAL CRUST.

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Chemical data can provide important evidence for processes operating during the generation of the earliest continental crust. Archaean continental crust can be broadly divided into low-grade (granite-greenstone) and high-grade (amphibolite- and granulite-facies) terrains. Rocks of tonalitic composition form a major component of both terrains. Low-grade terrains typically comprise greenstone belt sequences (ultramafic, mafic and silicic lavas and sediments) overlying a basement of tonalitic plutons and gneisses with a minor mafic component, with later intrusions of granites. The temporal relations between greenstone belt sequences and basement are often ambiguous. High-grade terrains comprise a bimodal suite of silicic (dominantly tonalitic) gneisses and mafic inclusions. Tonalitic gneisses greatly exceed mafic gneisses in volume, and generally comprise >75% of high-grade terrains. A variable proportion of metasedimentary material is present in high-grade terrains, but is often trivial in amount. Ultramafic-mafic-anorthosite complexes, often with associated metavolcanic amphibolites and metasediments, are a ubiquitous component of high-grade terrains, and are also commonly found in low-grade terrains. Intrusion of extensive dyke swarms appears to follow stabilisation of the newly formed continental crust.

Archaean deep (granulite-facies) crust is typically severely depleted in radioactive heat producing elements (1). This depletion appears to occur within 100-200 Myr of generation of juvenile crust (2), and could be the result of either partial melting or flushing of the lower crust by CO₂-rich fluids during granulite-facies metamorphism. Trace element data for contemporaneous Archaean amphibolite- and granulite-facies gneisses from the North Atlantic craton (3) and southern India (4) suggests that high-grade metamorphism is not accompanied by significant degrees of partial melting. The radioactive heat producing elements are strongly decoupled from other trace elements, and this is best accomplished via an active metamorphic fluid with a high CO₂/H₂O ratio. The source of the CO₂-rich fluids is debatable (5), but volumetric considerations would seem to require a mantle derivation. The middle and upper crust are the ultimate repositories of those elements removed from the lower crust during granulite-facies metamorphism: intermediate crustal levels undergo metasomatism and migmatisation, and granodioritic/granitic partial melts so formed may migrate into the upper crust. Thus the continental crust undergoes strong vertical chemical zonation at an early stage in its evolution.

Crustal growth in the Archaean occurs dominantly via the addition of tonalitic magmas to the continents. Archaean tonalites (both low- and high-grade) have many chemical characteristics in common with modern andesitic/tonalitic magmas associated with destructive plate margins (6). In particular, large ion lithophile (LIL) elements are enriched relative to light rare earth elements

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(LREE), while high field strength (HFS) elements (notably Nb and Ta) are strongly depleted relative to both the LREE and LIL elements. However, Archaean tonalites display a marked depletion in the heavy rare earth elements (HREE) which is not generally evident in otherwise compositionally similar modern rock types. Trace element constraints require a mafic source (amphibolite, garnet-amphibolite or eclogite) for the production of Archaean tonalites, and water appears to be an essential component during partial melting. Fusion of mafic rocks at appropriate F-T and under hydrous conditions yields liquids with the requisite major element chemistry, whilst stabilising residual minor phases (rutile, sphene, ilmenite) capable of retaining HFS elements.

Operation of some form of plate subduction in the Archaean provides a suitable tectonic setting for crustal generation. Rapid subduction of altered (wet), possibly komatiitic, ocean crust would provide a source for the production of voluminous tonalitic/trondhjemitic magmas. The higher geothermal regime in the Archaean may have allowed partial melting of the subducting slab to occur, rather than just dehydration of the slab as at the present. The juvenile tonalitic magmas so produced would be underplated to the base of the continental crust. The association of ultramafic-mafic-anorthosite complexes with metavolcanics and metasediments may represent slices of ocean crust which become tectonically emplaced into the deep levels of the continental crust during subduction (7), this occurring penecontemporaneously with underplating of tonalitic magmas. Mafic inclusions in the crust may represent deformed and disrupted syn-tectonic dykes. Production of the sub-continental lithosphere, and of komatiitic magmatism in greenstone belts, may also be related to subduction: the deeper levels of subducting ocean lithosphere might be the source of diapirs of depleted mantle which either underplate the continental crust (forming the tectosphere), or pierce/rupture the crust to produce ultramafic volcanism (8). Evidence from post-tectonic dyke swarms (9) indicates derivation from sub-continental mantle which is chemically heterogeneous, and which attained its chemical characteristics contemporaneously with generation of the overlying continental crust.

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DATING ARCHAEOAN ZIRCON BY ION MICROPROBE - NEW LIGHT ON
AN OLD PROBLEM

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One of the challenges of Archaeozoic geochronology is to find isotopic systems that preserve an indication of a rock's primary age in spite of the effects of later metamorphism. Zircon dating has been used widely with considerable success but not without difficulty, especially in polymetamorphic terrains. Zircons in such cases commonly are found to have lost radiogenic Pb, and despite fractionating the zircons¹ or abrading them to remove disturbed portions² it often is not possible to define a pattern of Pb loss from which the original age can confidently be inferred. The refinement of techniques to enable extremely small samples³, or even single crystals⁴, to be analysed has contributed greatly to solving the problem but even those techniques cannot resolve the micron-scale isotopic heterogeneities within single zircons in which much of their history is recorded. That can only be done by ion microprobe. We present here progress reports on studies of four Archaeozoic rocks, each of which illustrates the power and potential of ion microprobe analysis in solving problems of Archaeozoic geochronology.

Morton Gneiss. The gneisses of the Minnesota River Valley have long been recognized to be among the oldest rocks in the United States. However, because their isotopic systems have been disturbed by later metamorphisms and intrusions, there is uncertainty as to their actual age. The gneisses have been the subject of a number of geochronologic studies. Zircon U-Pb analyses by Catanzaro⁵ first showed the great antiquity of the rocks, giving a minimum age of 3250 Ma and an inferred age of 3500 Ma. There have been several zircon U-Pb studies since then^{6,7,8,9} but each, even the single crystal work⁸, has produced patterns of Pb loss from which the primary age cannot readily be deduced. Interpreted ages range from 3230 Ma^{7,8} to 3590 Ma^{6,9}. The Sm-Nd and Rb-Sr total-rock analyses of McCulloch¹⁰ support the older value, but other Rb-Sr work¹¹ has been interpreted to indicate ages even greater.

We studied a split from sample 673 analysed earlier by Goldich and Wooden⁹. It came from tonalitic Morton Gneiss 5 km NE of Delhi. The zircons are strongly zoned and at least five stages of growth can be seen, all, or only some of which may be present in any one crystal. A) Unzoned, euhedral, low-U (<500 ppm) cores overgrown by B) strikingly euhedrally zoned, moderate-U (1500 ppm) zircon, broken, then overgrown by C) unzoned, inclusion-rich, high-U (>2000 ppm) zircon, overgrown by D) unzoned or weakly euhedrally zoned low-U (<300 ppm) zircon, rarely overgrown by E) unzoned, moderate-U (800 ppm) euhedral rims. Type D commonly forms whole crystals. Plotted on a Concordia diagram the analyses for zircons type A to D all define, almost within analytical error, a single discordance line, with Concordia intercepts of 3535 ± 45 Ma and 1530 ± 200 Ma. Type E zircon is nearly concordant and unequivocally younger, 2585 ± 10 Ma based on $^{207}\text{Pb}/^{206}\text{Pb}$. The primary age of the bulk of the zircon is therefore probably 3535 ± 45 Ma, and is certainly greater than 3460 Ma. The 2585 ± 10 Ma age of the zircon rims is the same as the age of late intrusives in the Morton area⁹. It is very likely that the older zircons were also affected at that time so we presently attribute no age significance to the 1530 Ma intercept, except that it shows much of the zircon has suffered Pb loss at or after 1530 Ma. An excess of radiogenic Pb at one analysed spot suggests Pb movement may have occurred recently. Earlier conventional analyses of these zircons⁹ are readily explicable as mixtures of the components that the ion microprobe has identified.

Williams, I.S. *et al.*

Gneiss at Watersmeet. The Watersmeet Dome in northern Michigan is a second area in the United States where very early rocks have been identified. There also, later metamorphisms and intrusions have disturbed the older isotopic systems, making the determination of original ages difficult. Indications of Archaean ages were first found at Watersmeet in a Rb-Sr study by Sims and Peterman¹². Subsequent Rb-Sr work and zircon U-Pb analyses by Peterman and others¹³ showed that the gneisses were at least 3410 Ma, and possibly 3500 to 3800 Ma old. As in the case of the Morton Gneiss, the uncertainty in the age was due to the poor definition of the zircon Pb-loss pattern. Sm-Nd analyses confirmed the antiquity of the gneisses at Watersmeet. McCulloch and Wasserburg¹⁴ measured a total-rock model age of 3600 ± 40 Ma on one sample, Futa¹⁵ measured model ages of 3520 ± 70 Ma and 3919 ± 30 Ma (the latter was interpreted as being affected by metamorphism) on others. Recent zircon U-Pb work by Peterman and others¹⁶ has resulted in a revised zircon age estimate of 3562 ± 39 Ma.

The zircons studied by ion microprobe were from sample M83, a tonalitic biotite augen gneiss from 7 km NW of Watersmeet. Conventional zircon U-Pb analyses of M83 have been reported by Peterman and others^{13,16}. The ion microprobe analyses show a wide range in Pb/U and Pb isotopic composition in the zircons, even within single crystals. On a Concordia diagram the data do not define a single line. Areas of zircons with the lowest U (<400 ppm) have the highest $^{207}\text{Pb}/^{206}\text{Pb}$ and average 15% discordance. All have the same $^{207}\text{Pb}/^{206}\text{Pb}$ age, 3625 ± 13 Ma. Areas of zircons with the highest U (>800 ppm) mostly define a line with an upper Concordia intercept of 2645 ± 35 Ma. Analyses of intermediate-U zircon scatter in between. The original age of the gneiss is probably 3625 ± 13 Ma. The zircons are interpreted to have been partially or totally reset at 2645 ± 35 Ma and all the zircons have again lost radiogenic Pb at least as recently as 300 Ma. The 2645 ± 35 Ma Pb loss event corresponds to the time of intrusion of younger dykes and granites in the Watersmeet area^{13,16}. Mixtures of zircons of these compositions account for the scatter observed in the conventional zircon analyses.

Gneiss at Mount Sones. The Napier Complex of Enderby Land, Eastern Antarctica, is a granulite grade terrain that despite having undergone at least three major episodes of deformation and metamorphism still preserves some isotopic evidence of great antiquity^{17,18,19,20,21,22}. One sample from the Fyfe Hills has been shown by ion microprobe analysis to contain a minor zircon component at least 3800 Ma old¹⁸. A much greater concentration of old zircon has now been found by ion microprobe in a granodioritic orthogneiss from Mount Sones. The U-Pb systems in the zircons are unusually complex. Most of the grains are brown and markedly zoned, a few are pale and structureless. Analyses of the pale zircons mostly plot near the Concordia at ~ 2400 Ma, recording the metamorphism at 2456 ± 7 Ma identified by Black and others¹⁹. The brown zircons show a much greater range in composition which can be related to the three visible stages of crystal growth: A) cores overgrown by B) euhedrally-zoned zircon, overgrown by C) relatively structureless rims. Analyses of type C zircon mostly plot on a short discordance line that intersects the Concordia at ~ 3000 Ma, recording the granulite facies metamorphism at 3072 ± 34 Ma²⁰. The analyses of zircons type A and B are very widely dispersed; some are reversely discordant indicating an excess of radiogenic Pb. They scatter about a line with Concordia intercepts of ~ 2400 Ma and ~ 3750 Ma. If the type A zircons are regressed separately, their upper intercept is ~ 3850 Ma. At one 35 μm spot on a type A zircon a large and variable excess of radiogenic Pb was found. A correlated change in the Pb/U and $^{207}\text{Pb}/^{206}\text{Pb}$ indicates that Pb was gained at the spot ~ 2400 Ma ago, evidently in response to the same metamorphism that in most other zircons caused Pb loss. $^{207}\text{Pb}/^{206}\text{Pb}$ ages up to 4200 Ma were

Williams I.S. *et al.*

measured on the spot, showing that $^{207}\text{Pb}/^{206}\text{Pb}$ ages measured on very small zircon samples need not necessarily be minimum ages.

Quartzite at Mount Narryer. One of the aims of Archaean geochronology is to determine when the earliest terrestrial crust formed. An ideal place to look for surviving remnants of that crust is in the earliest sediments. Dating zircons in sediments by conventional methods is both difficult and tedious for in most cases the zircon population is a mixture of zircons from many rocks of different ages. The ion microprobe however, with its ability to make rapid analyses of single crystals, is ideally suited to the task. During a reconnaissance survey of zircons in Archaean metasediments from a variety of localities in Western Australia, zircons have been identified that are up to 4200 Ma old²³. The old zircons occur in a quartzite at Mount Narryer, 250 km NE of Geraldton. They comprise about 5% of the zircon in the rock, most of the remainder being zircon formed at ~ 3750 Ma or ~ 3500 Ma. The analyses of the old zircons are very nearly concordant. The possibility that the measured age is nevertheless an artifact of inherited radiogenic Pb was tested by analysing one crystal seven times. Despite a range of a factor of 1.5 in the U content of different areas of the grain, Pb/U and $^{207}\text{Pb}/^{206}\text{Pb}$ were virtually constant, arguing strongly that excess radiogenic Pb is not present. The existence of the old grains shows that silica-saturated rocks were present on the Earth's surface as early as 4200 Ma and that they were a significant fraction of the surface exposure at the time the original sand was being deposited, probably sometime between 3300 Ma and 3500 Ma. This evidence that some early rocks did survive the pre-3900 Ma meteorite bombardment of the Earth gives hope that intact segments of that crust may yet be found.

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